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Boris S. Sokolov

Academician, Russian Academy of Sciences
counsellor of the Presidium of the Russian Academy of Sciences

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The Founder of the Russian Geological School

On the Occasion of the 150th Anniversary of A.P. Karpinsky, the First Elected President of the Russian Academy of Sciences

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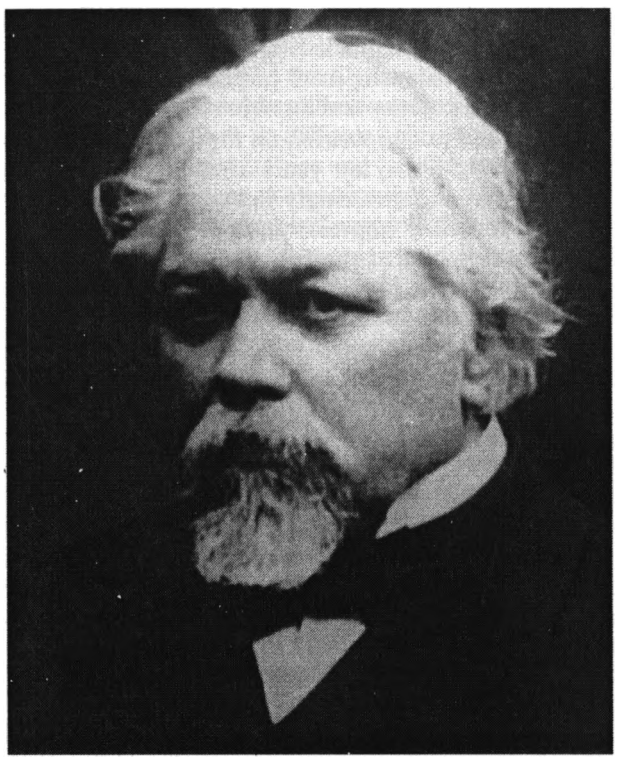

Aleksandr Petrovich Karpinsky was born on December 26, 1846 (January 7, 1847 by the Gregorian calendar) in a settlement near the Bogoslovskii Mining Plant of the Verkhoturskii District in the Urals; it is a town now called Karpinsk. Karpinsky came from a family of hereditary mining engineers. His grandfather and father were both educated at the Mining Military School in St. Petersburg, at that time a closed, semi-military institution. Following the tradition of the family, the twelve-year-old Karpinsky entered the Institute of Mining Engineers of that school in 1858 and graduated in June 1866 with a minor gold medal as a mining engineer at the rank of lieutenant. Incidentally, in 1866, this institution was renamed as the St. Petersburg Mining Institute. It was famous for a long history of extensive and various educational courses, a splendid group of professors, and it was ranked with the best higher educational institutions in Russia.

The traditions of the Karpinsky family, which for generations lived in the mining district of the Urals, seemed to determine the scope of activity of Aleksandr Petrovich, and he actually worked for about a year and a half at mining fields and deposits in the Urals. But his future career took another course. He was soon back at the Mining Institute as an adjunct preparing himself for a professorship. In January 1868, he started his never-ending, extremely intensive, and diverse scientific, pedagogical, engineering–geological, organizational, and public activity. He amazed his contemporaries with his productivity and his truly Lomonosovian spirit of deep insight into numerous fields of knowledge, even those that were far beyond his immediate scientific intersts.

The contribution of Karpinsky to the development of geology and geological education in Russia and to the restoration and progress of the Russian Academy of Sciences is of unique importance. These two aspects of his long and distinguished activity were reflected in both Russian and world literature, in journals and various publications on the activities and history of the Academy itself, in the development of the geological service in this country, which covers vast areas on two continents. It may seem strange that, except for a few reviews of his activity (Lichkov, 1946; Bayan, 1955; Chernousov, 1962; Kumok, 1978), a special collection of papers, and the addresses published in memory of this great deceased scientist and organizer (a special issue of *Vestnik Akademii Nauk SSSR*, 1936, no. 7 and collection of articles entitled *A.P. Karpinsky,* 1937), only a single book has so far been published that can claim to be a sufficiently complete scientific biography of Karpinsky (Romanovskii, 1981). I am inclined to attribute this phenomenon to the fact that, though most of his numerous (more than 500) published works were short, about a hundred of them were so complete and impregnated with original ideas that they were immediately and extensively absorbed by the scientific community and became generally acknowledged without particular monographic or commentary analyses. Of considerable importance was his professorial activity at the Mining Institute, which he began when he was 22 years old and continued for 30 years. His ideas, always based on profound studies of factual material and vast geological experience, were presented in such an unimposing, lucid style that they were recognized as

true knowledge in its most perfect form. The stock of ideas of the teacher was easily absorbed by his disciples searching for their own way in science. Members of the Russian Academy of Sciences F.N. Chernyshev, V.A. Obruchev, E.S. Fedorov, A.A. Borisyak, N.N. Yakovlev, later D.V. Nalivkin, and many other geologists were his students.

In 1882, the establishment of the Geological Committee, now the Karpinsky All-Russia Institute of Geology (VSEGEI) in St. Petersburg, was a great stimulus to geological research and the practical training of future geologists in Russia. Actually, this state geological institution of national importance was organized on the initiative of geologists and professors at the Mining Institute and St. Petersburg University who were closely associated with the Academy of Sciences. The Committee was headed by Academician G.P. Gel'mersen, the eighty-year-old tutor of Karpinsky and the author of the first "General Map of Mountain Formations in the European Part of Russia." But the life and soul of the Geological Committee was Karpinsky; in 1884 he was nominated as director and officially held this post for more than twenty years, and in fact never lost contact with the Committee as its honorary director until 1936. A remarkable feature: to this day, the door of his spaceous study in the building of the Geological Committee, now VSEGEI, in St. Petersburg has a modest plate inscribed simply "A.P. Karpinsky".

Karpinsky had excellent knowledge of the 50 years experience of geological survey institutions in European countries (in Great Britain since 1832), and he held that the main task of the Geological Committee in Russia was to conduct regular geological surveys and geological cartography as a point to supply the economy of the country with mineral resources. He also clearly defined this kind of geological activity as versatile scientific research based on deep knowledge of the stratigraphy, paleontology, and petrography of volcanic and sedimentary rocks, of general geology and tectonics, and of mineral resources. For many decades this approach was the basic style for geological works of the Geological Committee of Russia and other institutions, which gave the Russian geological school a reputation as one of the best-known in the world. The inevitable differentiation of geosciences has opened new vistas for specialization of geological knowledge, but it did not alter the principal importance of geological surveying as the scientific research producing basic materials for theoretical generalizations in geology and effectual application of geological data in practice. Today we may note with regret a falling away from the principles inspired by Karpinsky and the detrimental influence of federal red tape, regionalization, and the extreme specialization of the geological service to the detriment of geological surveying.

Academician N.S. Schatsky has written a series of brilliant articles dedicated to Karpinsky and incorporated in volume IV of his "Selected Works" (Schatsky, 1965, pp. 233–273). In one of the papers that later reappeared in the well-known "Portraits of Geologists" (Schatsky and Yanshin, 1986, pp. 93–110) he writes:

> Among Russian geologists of this and the last centuries we fail to find a scientist equal to Karpinsky in the versatile character of his scientific interests, in the scope of his scientific creative power, and in the scrupulous elaboration of scientific problems. Karpinsky is regarded as the progenitor and founder of Russian geology. But the significance of his geological activity is much more extensive. Karpinsky was the first not only in the task of organizing geological surveying in Russia and of establishing the largest and the best geological school in the Soviet Union. He is also our first theoretician in geology, an unequalled master who greatly advanced the development of geological sciences in his home country and far abroad.

These words were written on the 100th anniversary of Karpinsky's birth, ten years after the death of this great scientist and amazingly harmonious personality. He left an indelible impression on the minds and spirit of the people who had the opportunity to be his pupils and to work with him. Centuries may pass without obscuring his individuality.

Karpinsky was always persistent in pursuing the comprehensive approach to geological research, in the responsibility he assumed in their organization, and in the principles elaborated for geological education. He was always delighted by the achievements of his colleagues and pupils, but modest in estimating his own scientific discoveries and results, no matter how distinguished. Encouraged by Professor N.P. Barbot de Marni, he chose to study geology rather than mining engineering; his first interest was petrography and mineral deposits, both ore and sedimentary, mainly coal. He was the first to introduce the microscopic method in rock study and to elaborate an original classification of rocks. He obtained his most spectacular results, however, in his study of stratigraphy, paleontology, historical geology, and tectonics. In this quest, he approached the problem of the morphology, distribution, and structure of continents from an entirely different angle, which we would now classify as one of the principles of "global plate tectonics." His knowledge of the world literature and the diversity of his geological intersts were extraordinary.

Once, half in jest, D.V. Nalivkin said that any researcher becomes a true geologist only after he has worked in the Urals, so instructive and complicated are the structure and geological history of this unique mountain belt. In this sense, Karpinsky was indeed a Uralian geologist not only by birth, but by his geological experience, which enabled him not only to produce admirable cartographic works, but also to make many discoveries, particularly in the Upper Paleozoic. Among the latter is the fundamental substantiation of the Artinskian Stage of the Permian System (1874), the deposits of which, after Pallas, Gregory, and Murchison, were treated ambiguously as "Permo–Carboniferous." Karpinsky had a special aptitude for unravelling

such difficult stratigraphic notions that enabled him to elaborate a general conception of the "transition layers" in stratigraphic classification, a concept still provoking heated debates at numerous theoretical discussions between stratigraphers.

Karpinsky was interested in both the intricately dislocated folded belt of the Urals and the geologically different gigantic territory of the European part of Russia (the East European or Russian platform). The platform is a kind of "table country" with quite another type of tectonic movements, which he termed "crustal oscillations," manifested in a form of "gentle syncline and anticline." For more than three decades (1880–1919) Karpinsky dedicated his research to the geological history of the European part of Russia framed by the Urals and the Caucasus. He considered the last ranges not only as the bounding folded frame, but as structures that at different time periods differently influenced the tectonics and relief formation of the platforms. In the course of geological history, the gradual deformations of the platform relief with their own laws regulated the coast line migration in ancient seas (transgressions and regressions). In 1883, at the beginning of his research in the South European part of Russia, geologically the most active area, he established the zone of specific dislocations termed "incipient ridge bands," which was later called by E. Suess "the Karpinsky lines" (the Donets basin was a part of that band). Most important from the view point of historical–tectonic analysis were his generalizations on the character of crustal oscillations in the European part of Russia.

The following publications, though not large, should be mentioned here as works that ushered in an entirely new epoch in the methods of study of ancient platforms of the Earth: "A Review of the Physical–Geographical Conditions in the European Part of Russia in Past Geological Periods" (1887), "The General Character of Crustal Oscillations Within the European Part of Russia" (1894), "On the Tectonics of the European Part of Russia" (1919); these papers were also published later, in 1947, under the general title "Essays on the Geological Past of the European Part of Russia." Under the influence of these publications, the international map of Europe was improved, the Russian part of which was edited by a team where Karpinsky had the leading role from the beginning. The first of the enumerated papers was of fundamental importance for the application of the paleogeographic method when reconstructing the geological past, though Karpinsky did not use the term paleogeography. In North America, in 1910, similar work was done by Ch. Schuchert, but as N.S. Schatsky noted in 1947, his results "did not evoke such important tectonic conclusions as those made in the work of Karpinsky." No less fundamental was his elaboration of the role of oscillatory crustal movements in the formation of specific plicative structures of the platform and in the dynamics of the coastal lines of ancient seas. Almost simultaneously with Karpinsky, F.Yu. Levinson-Lessing wrote in 1893 about "the oscillations of the lithosphere." The famous French geologist E. Haug, in 1913, in his discussion of ideas of many European geologists and geomorphologists on the history of the Mediterranean coastline, which they considered eustatic, preferred the views of Russian geologists to those of the "French stratigraphic school," though in this period (the 1880s–1890s) the difference between the orogenic and epeirogenic movements was still obscure.

During his long life and intensive activity, Karpinsky almost invariably pursued several courses of investigation. His various interests in geology and natural sciences in general, to which we should add the humanitarian sphere and music, moved his attention from one topic to the other and involved him in innumerable scientific–organizational and social affairs where he was always scrupulously punctual, conscientious, and energetic. But there was one sphere of science that particularly attracted him—paleontology and its impact on stratigraphy and geological cartography, paleobiological and evolutionary problems, and the unravelling of the nature of enigmatic organisms. As a conscientious stratigrapher with a sound paleontological and petrographical (we should now say lithological) schooling, he became a member of the Russian Committee on the unification of geological patterns in geological maps, which was initiated by the International Geological Congress (IGC) in 1878 in France. In 1881, at the Second Session of IGC in Bologna, Italy, he was awarded the prize of the Congress for his "experience in the systematic unification of graphic representations in geology." His classification and nomenclature that represented subdivisions of sedimentary crustal layers have remained almost intact to this day, though the general and particular problems in stratigraphy itself are still extremely numerous and keen.

In 1962, A.A. Borisyak, depicting Karpinsky as a paleontologist, said that he was "not only a geologist, but a biologist as well. And it was as a biologist that he considered paleontologic remains when he first introduced in our literature the ontogenetic method for the study of ammonites, or the histological method in the study of fishes. Consequently, his paleontological works are of immense importance. Extremely conscientious research allowed Karpinsky to establish the phylogenetic relations between ammonoids that he studied, in other words, to draw their genealogical tree." For this achievement the Academy of Sciences of France awarded him the Cuvier Prize. His series of papers (about ten) on the very mysterious *Helicoprion* fish is absolutely unique, like those on the Late Paleozoic cladoselachians in general. Of quite exceptional interest is his monograph on the scrupulous study of oogonia of the Devonian charophytas that could have brought honors to any paleobotanist. After the publication of these works, a stream of strange fossils began to reach Karpinsky from many countries; he studied such fossils for many years and formed a special trend in paleontology, the study of problematics.

Already in the last century, Karpinsky was the most famous Russian geologist in Western Europe and America. Numerous academies of sciences and scientific societies (and not only geological) elected him their honorary member, or corresponding member and awarded him honorary medals and other awards. He was an active participant of many geological congresses, an active member of various commissions, the first recipient of the Spendiarov Prize of the IGC in 1900, the vice-president of the VI IGC session in 1894 in Zürich, the president of the VII IGC session in 1897, which was convened in Russia for the first time. In August 1935 he was elected honorary chairman of the organizing committee of the XVII IGC session held in Moscow in 1937, but Aleksandr Petrovich could not attend it. He died on July 15, 1936, near the age of ninety.

The activity of Karpinsky outlined above was invariably combined with his work as professor of the St. Petersburg Mining Institute, and as the head of the Geological Committee of Russia, and also with his duties as academician and president of the Russian Academy of Sciences, which in 1934 was transferred from Leningrad to Moscow. The attitude of Karpinsky towards the Academy was almost reverential; he was very grateful to be its member and spared no efforts to maintain the honorable reputation of the Academy, which in the 1930s experienced considerable difficulties. He became a member of the Academy as early as in 1886 after being an adjunct at the physico-mathematical department (Geology); then in 1889 he was unanimously nominated extraordinary academician and in 1896 ordinary academician. These were the years of his intensive work as director of the Geological Committee of Russia, and he officially left this position only in the spring of 1903 owing to his increasing duties at the Academy; however, he was the honorary director of the Committee until his death. Since the establishment of the Russian Academy in 1724–1725, Karpinsky was the sixth geologist to hold the rank of academician, and the Academy rated high his fields of activity, his personality, his authority, and worldwide fame as a scientist. In 1886, at the ceremonial General Meeting of the Academy, he delivered a speech "On the Physico-Geographical Conditions of the European Part of Russia in the Past Geological Periods". A year later it was published in the papers of the Academy of Sciences. The exposition and conclusions of the author were so full of new conceptions that the publication was immediately acknowledged as part of the "golden fund" of world geological literature. Karpinsky installed the geological sciences in the Academy by showing their new and extremely important relations with science as a whole. It was an achievement that has forever secured a significant position for geosciences in the structure of the Russian Academy of Sciences. The principal role of Karpinsky here can hardly be overestimated.

By historical tradition, the presidents of the Russian Academy of Sciences, which was then called Imperial, were always nominated by edict of the Emperor. But the last president, the Grand Duke Konstantin Konstantinovich Romanov, died on June 2, 1915, and the vice-president died soon after. The war and social havoc greatly impaired the functions of the Academy, and Nicholas II gave his consent to entrust Karpinsky with the temporary execution of the duties of Vice-President from May 15, 1916, which actually meant the general administration of the Academy. In 1917, after the February Revolution, the Provisional Government, for the first time in Russia, issued an edict that the president and the vice-president of the Academy were to be elected from among the ordinary academicians. The nomination of Karpinsky to the post of president was an obvious fact to all. On May 15, 1917, the Extraordinary General Meeting of the Academy unanimously elected him the president of the Academy. The adjective "Imperial" was later abolished by a special resolution of the Provisional Government stating that "the former Imperial Academy of Sciences should henceforth be named the Russian Academy of Sciences." (Romanovskii, 1981).

Academician Aleksandr Petrovich Karpinsky, therefore, was President of the Russian Academy of Sciences from before the events of October 1917, and from then he was continuously its president for a full nineteen years. He died on July 15, 1936, at his country house near Moscow, and his remains are entombed near the Kremlin wall in Red Square.

With this short article that opens the present issue, which is dedicated to the 150th Anniversary of Academician A.P. Karpinsky, one of the famous geologists of the 19th and 20th centuries and the first elected President of the Russian Academy of Sciences, the editorial board of the journal intends to initiate, in the course of 1997, the publication of a series of papers, which will show the evolution of the scientific heritage of this outstanding scientist, will reflect the development of his ideas, and will be related to his exceedingly prodigious geological interests.

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[1] Includes the articles: Aleksandr Petrovich Karpinsky (1958); A.P. Karpinsky as a Scientist (on the Occasion of the 100th Anniversary of His Birth) (1948); Works of Karpnskii on Tectonics of the East European (Russian) Platform (1947); The Conceptions of Karpinsky about the Causes of Faulting (1965); On the Influence of N.P. Barbot de Marni on A.P. Karpinsky as a Geologist.

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The Second Guadalupian Symposium and Problems of Permian Stratigraphy

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Abstract—The article presents information on the Second Guadalupian Symposium in April 1996 and a meeting of the International Subcommission on Permian Stratigraphy held in its framework. The main subjects for discussion both at the Symposium and the Subcommission meeting were the Standard Stratigraphic Scale for the Permian System and, particularly, the problem of introducing the Guadalupian Series into the scale. The discussions resulted in the recommendation to subdivide the Permian System into three series: the Cisuralian Series with Asselian, Sakmarian, Artinskian, and Kungurian stages, the Guadalupian Series with Roadian, Wordian, and Capitanian stages, and the Lopingian Series with Wuchiapingian and Changhsingian stages.

Key words: *The Permian, stratigraphy, standard scale, the Guadalupian Symposium, International Subcommission on Permian Stratigraphy*

The Second Guadalupian Symposium and a meeting of the International Subcommission on Permian Stratigraphy (SCPS) were held in Alpine, western Texas, United States, in April 10–13, 1996. The symposium was organized by the United States Geological Survey, Sul Ross State University, and the SCPS. It was attended mainly by specialists from the United States. Among the invited scientists, there were experts from Russia (T.A. Grunt, V.I. Davydov, G.V. Kotlyar, E.Ya. Leven, and T.B. Leonova), China (Jin Yugan), Canada (T. Tozer, M. Orchard), Germany (M. Menning), Hungary (H. Kozur), and Switzerland (A. Baud). The problems of substantiation of the Guadalupian Series and related revision of the Standard Permian Scale were in the center of the program of the symposium. Eighteen reports and a two-day excursion to the Permian sections of the Glass and Guadalupe mountains were devoted to these topics. They were also discussed at the SCPS meeting, where an important revision of the Permian scale was suggested. The present article will summarize the discussions and make some comments on these decisions. However, previous studies of the discussed problems should be dwelt upon briefly.

At the First Guadalupian Symposium held in March 1991, it was decided to propose the Guadalupian Series as the third one in the Permian System and to chose a section of the Delaware Basin of western Texas, in the United States, for its stratotype. The proposal was officially made by B. Glenister from the University of Iowa at the congress "The Permian System of the Globe," which took place in the town of Perm' (Russia) in August 1991 (Glenister *et al.*, 1992). At the SCPS meeting held during the congress, an idea was advanced to divide the Permian System into two subsystems (or series), each comprising two series (or subseries) (Leven, 1992). In this case, the Guadalupian would correspond to the lower series (subseries) of the upper subsystem (series) (table). This idea was presented in the "Operational Scheme of Permian Chronostratigraphy," which was discussed at the International Symposium on Permian Stratigraphy, Environments and Resources in the town of Guiyang, China, in August 1994 (Jin Yugan *et al.*, 1994a) (table). According to this scheme, the Permian System was divided into four series: Uralian, Chihsian (Cathedralian), Guadalupian, and Lopingian. The scheme elicited lively discussion, but it has not been accepted. The weak point of the scheme was the position of the Kubergandian Stage below the Roadian Stage, although the ammonoid assemblages characteristic of these stages were known to be similar. Opinions diverged with regard to conodont identifications and their stratigraphic succession in the upper Guadalupian boundary interval.

A revised version of the Operational Scheme was published immediately after the symposium (Jin Yugan *et al.*, 1994b) (table). Both versions were discussed in a special article (Leven *et al.*, 1995) and in some publications of *Permophiles* (no. 25, 1994). Therefore, there is no need to consider them again here. However, it should be noted that the first version has been considerably changed. Obviously, the changes were made not because of new data obtained (there was

Variant divisions of the Permyan System into series and stages

Glenister, 1992	Leven, 1992		Jin Yugan *et al.*, 1994		Permophiles, 1994			Alpine, 1996	
Upper series	Upper subsystem	Upper series	Lopingian	Changhsingian	Upper series	Lopingian	Changhsingian	Lopingian	Changhsingian
				Dzhulfian (Wuchispingian)			Wuchiapingian		Wuchiapingian
Middle series (Guadalupian)		Lower series	Guadalupian	Capitanian		Guadalupian	Capitanian	Guadalupian	Capitanian
				Wordian			Wordian		Wordian
				Roadian			Roadian		Roadian
Lower series	Lower subsystem	Upper series	Chihsian (Cathedralian)	Kubergandinian	Lower series	Leonardian	Cathedralian	Cisuralian	Kungurian
				Bolorian			Hessian		Artinskian
		Lower series	Uralian	Artinskian		Cisuralian	Artinskian		
				Sakmarian			Sakmarian		Sakmarian
				Asselian			Asselian		Asselian

no time for this), but due to pressure from critics. Therefore, the second version was no more convincing than the first. Additional investigations were needed particularly to define more accurate positions of the boundaries between the suggested series and to show their traceability.

The following steps were taken for the year and a half before the Second Guadalupian Symposium.

(1) H. Kozur (Hungary) and Wang Zhihao (China) reidentified the conodonts from the Luodian section of southern China (Kozur's pers. communic.). It was ascertained that the typical Roadian, i.e., early Guadalupian conodonts appeared not within the Murgabian Stage, as shown in the first Guiyang version of the scheme, but in Bed 25b (*International ...*, 1994) referrable, in Leven's opinion, to the lower Kubergandinian. This completely agrees with the ammonoid data and allows correlation between the lower boundary of the Roadian Stage and, hence, of the Guadalupian Series of the North American scale and the base of the Kubergandinian Stage of the Tethyan scale.

(2) Identifications and stratigraphic succession of conodonts from the Guadalupian–Lopingian boundary interval were revised (Kozur, 1995). The conodonts *Clarkina crofti* and *C. postbitteri* characteristic of the base of the Lopingian Group (and that of the Wuchiapingian Stage) of southern China appeared to occur at the top of the Guadalupe Group of the Delaware basin (within a few upper centimeters of the Lamar Limestones and Altuda Formation) (Kozur, 1996). This allows us to draw the Guadalupian–Lopingian boundary at the base of the *C. postbittei–C. crofti* Zone of the Wachiapingian Stage. The scope of this stage exceeds the Dzhulfian Stage of Transcaucasia, because the basal *C. leveni* Zone of the latter stage occurs inside the Wuchiapingian Stage.

(3) Changes in various marine fossils at the end of the Sakmarian–beginning of the Artinskian ages were analyzed. It was revealed that they had a subglobal character and were related to the alteration in transgressive–regressive cycles, which were probably accompanied by climatic changes (Leven *et al.*, 1996). This is a strong argument for subdividing the Lower Permian into two independent series.

(4) The Permian sections of the Glass and Guadalupe mountains were examined again and the conodont-based boundaries of the Cathedralian, Roadian, Wordian, and Capitanian stages were defined more accurately.

The newly obtained data undoubtely supports the recognition in the Guadalupian Series in the Permian System. Previously, the Lower–Upper Permian boundary was also drawn at the bases of the Roadian Stage of the North American scale and the Kubergandinian Stage of the Tethyan scale. The data confirming the isochroneity of this boundary in the Tethyan and North American regions demonstrates its significance as a boundary between series. This boundary marks a significant renewal of fossil assemblages caused by the subglobal transgression well pronounced both in the Tethyan area and North America (Leven, 1993). Events that occurred at the boundary between the Guadalupian and Lopingian were no less important. The subglobal regression caused the extinction of many marine organisms abundant in the Guadalupian time, e.g., larger fusulinids of Schwagerinida and Neoschwagerinida orders, goniatites, agoniatites, many groups of corals, etc. (Kotlyar *et al.*, 1983, 1989; Leven, 1993). The Lopingian deposits of most sections are very distinct from

the Guadalupian sediments both in lithological composition and fossil content. The correlation between the base of the Lopingian Group of southern China and the top of the Guadalupe Group of North America showed that these events occurred simultaneously over the globe and the boundary between these series can be traced in marine sections of different continents.

If recent data allow the recognition of the Guadalupian Series and its boundaries, the same cannot be said of the subdivision of the Lower Permian into two independent Cisuralian and Leonardian series (subseries), as suggested in the second version of the "Operational Scheme of Permian Chronostratigraphy" (Jin Yugan *et al.*, 1994b). In principle, the mid-Early Permian events are comparable in significance with the events of the early and late Guadalupian Epoch. Therefore, the subdivision of the Lower Permian into two series or subseries seems to be logical and well substantiated. However, this subdivision is impossible until the problem of the boundary between these series has been solved. According to the available data, the signs of the above-mentioned events are not confined to the traditional boundary between the Sakmarian and Artinskian stages, but they are recorded inside the latter stage (Leven, 1994; Leven *et al.*, 1996). Therefore, the problem of the boundary between the suggested series can be solved through the reduction of the Artinskian Stage with the inclusion of its lower part into the Sakmarian Stage. This is why it is absolutely unacceptable to draw the Cisuralian–Leonardian boundary between the Artinskian and Hessian stages, as is suggested in the second version of the Operational Scheme. In fact, Ch. Ross, the author of the Hessian Stage, correlated it with the Artinskian and the underlying Lenoxian Stage with the Sakmarian (Ross and Ross, 1987). In the suggested version, the major part of the Hess Formation, which was identified by Ch. Ross as the Hessian Stage, is included in the Cathedralian Stage. The Hessian Stage is retained with a very reduced scope (the lower 15 m of the Hess Formation in the stratotype section). The conodonts found in this stage are still of Artinskian age (the *N. pequopensis* Zone). In the upper part of the underlying Lenox Hills Formation, there are only conodonts of the lower Artinskian *S. whitei–M. bisseli* Zone. Thus, the Cisuralian–Leonardian boundary lies not at the top of the Artinskian Stage, as shown in the scheme discussed, but within this stage. This position is unacceptable unless one revises the scope of the Artinskian Stage.

Before the symposium, B. Wardlaw (newly elected SCPS Chairman), H. Kozur, G.V. Kotlyar, E.Ya. Leven (SCPS members) and T.B. Leonova met in Washington to discuss the problems of Permian stratigraphy. Because an accurate definition of the Cisuralian–Leonardian boundary is still impossible and the beds recognized as the Hessian Stage are of Artinskian age, the Russian scientists proposed identifing the pre-Guadalupian part of the scale as a single series comprising the Asselian, Sakmarian, Artinskian, and Kungurian stages; the latter stage would then replace the Cathedralian Stage. The lower Kungurian boundary was suggested to be lowered to the base of the *N. pnevi* Zone. In this case, the Kungurian, Cathedralian, and Bolorian stages of the Uralian, North American and Tethyan scales respectively will appear to be equal in scope, but the first stage will have a historical priority. There were doubts whether the Kungurian Stage could be officially accepted, because its lower boundary in the Uralian sections did not meet international requirements, and, if this were the case, whether it would be possible to choose the international standard in the North American sections, having retained the name "Kungurian." It was decided to apply to the International Commission on Stratigraphy to work out these questions.

As for the Guadalupian and Lopingian series, there were no divergent opinions on their recognition within the suggested boundaries. However, Kozur spoke in favor of retaining the Dzhulfian Stage in the Lopingian Series and establishing between this stage and the Guadalupian Series a new Laibinian Stage in the scope of the conodont *C. postbitteri–C. crofti* and *C. niuzhuangensis* zones.

In Washington, it took little time to reach agreement on all the questions discussed. It was decided to submit to the SCPS meeting the three-member rather than the four-member scale of the Permian System. Later, this decision was supported by Glenister and Jin Yugan.

The SCPS met in Alpine after the two-day work of the symposium, at which 18 reports on general and particular questions were delivered (*Second* ..., 1996). In his opening speech, B. Glenister indicated the urgent need for a standard Permian stratigraphic scale based on marine sections and outlined its main principles. H. Kozur gave a comprehensive report dealing with the elaboration of this scale. In his opinion, the Permian System should be divided into three series. The lower series comprises the Asselian, Sakmarian, Artinskian, and Kungurian stages with type sections in the Uralian region. The middle series includes the Roadian, Wordian, and Capitanian stages with stratotypes in the Delaware basin of the United States. The upper series consists of the Laibinian, Dzhulfian, and Changhsingian stages with stratotypes in southern China and Transcaucasia. Boundaries between the Asselian, Sakmarian, and Artinskian stages should be more accurately defined. The Asselian–Sakmarian boundary is suggested to be drawn at the appearance of *Sweetognathus merrilli* and *Streptognathodus barskovi* (*S. postfusus*) s. s. In this case, it lies slightly lower than the traditional boundary, i.e., within the upper zone of the Asselian Stage. The Sakmarian–Artinskian boundary is suggested to be defined by the appearance of *Streptognathodus florensis*, in the event that this species will be found near the traditionally accepted boundary in the Uralian area.

G.V. Kotlyar reported on the principal correlation levels of the Permian System such as the bases of the Roadian ("Kubergandinian") and Dzhulfian stages with the latter being taken as an equivalent of the Wuchiapingian Stage of the Chinese scale. Problems of correlation were also discussed in the joint report of Kotlyar, Yu.D. Zakharov, and G.P. Pronina. In their opinion, the Roadian Stage of the North American scale corresponds to the Kubergandinian Stage and the lower zone of the Murgabian Stage of the Tethyan scale, while the Wordian Stage corresponds to the *N. crofticulifera–A. schencki* Zone of the Murgabian Stage and to the lower half of the Midian Stage, and, finally, the Capitanian Stage is equal to the upper half of the Midian Stage.

In his very interesting report, S. Lucas (United States) analysed the changes in Permian tetrapod assemblages throughout the world. It was found out that two significant crises in tetrapod development occurred in the terminal Artinskian and terminal Guadalupian ages. The conclusion was reached that, according to the distribution of tetrapods, it would be more natural to divide the Permian System into three series. There was a question about the accuracy of the dating of the first crisis. Lucas answered that the crisis was dated by conodonts from the marine limestones wedging the tetrapod-bearing continental deposits of the Robledo Mountains, New Mexico, United States. However, according to Kozur and Le Mone (1995) and Wardlaw (pers. communic.), both of whom identified the conodonts, these fossils only indicate the late Sakmarian–early Artinskian age. If so, the greatest crisis in the tetrapod development almost coincides in time with the above-mentioned Artinskian crisis of marine biota (Leven *et al.*, 1996). This supports the subdivision of the Lower Permian into two independent units.

T.A. Grunt analyzed the development of the Late Permian brachiopod assemblages in the equatorial area and indicated a strong provincialism of the North American (Midcontinent) assemblages. For this reason, the North American stages can hardly be used as standards of the international scale.

The possibility of correlating the Upper Permian and Lower Triassic deposits by means of paleomagnetic records was demonstrated in the report of M.B. Steiner (United States). It was shown that the lower part of the Chhidru Formation of the Salt Range is comparable with the lower part of the Changhsingian Stage and the upper part of the Wuchiapingian Stage of southern China, while the Wargal Formation can be correlated with the major part of the upper substage of the Wuchiapingian Stage. The upper Tatarian Substage corresponds to the Wargal Formation and the lower part of the Chhidru Formation in the Salt Range, to the Wuchiapingian Stage of southern China, and to the Capitanian Formation and the lower part of the Ochoa Formation in North America.

Among the general reports, the report of Jin Yugan (China) on recognition and correlation of the Lopingian Series within the Pangea and the report of M. Orchard (Canada) on the conodonts from the Permian–Triassic boundary deposits should be mentioned. Several reports analyzed various aspects of the Permian stratigraphy in the Delaware basin and characterized lithologic, paleontologic and paleomagnetic features of the constituent stages.

After the meetings the participants had a two-day excursions to the Glass Mountains and to the southeastern slopes of the Guadalupian Mountains. In the first area, participants examined the Hessian–Cathedralian boundary, which was suggested to be drawn at the appearance of *N. pnevi*. As mentioned above, this boundary lies 15 m above the base of the Hess Formation and is defined by conodonts only; fusulinid assemblages occurring under and above this boundary are almost the same, and ammonoids known from the middle part of the Hess Formation in the east (the Hess Fossil Bed) still show the Artinskian aspect.

On the second day, the suggested stratotypes of the conodont-based stage boundaries of the Guadalupe Group were shown. The lower boundary of the Roadian Stage, i.e., of the whole Guadalupe Group, is defined by the replacement of species *M. idahoensis* and by *J. nankingensis* within the continuous clayey limestones of the El Centro Beds in the middle part of the Cutoff Formation. The lower Wordian boundary is recognized at the appearance level of *J. aserrata* in the upper limestones of the Getaway Beds in the Cherry Canyon Formation exposed at the Guadalupe pass. Finally, the lower Capitanian boundary is fixed at the replacement level of species *J. aserrata* by *J. postserrata* within a thin stratum of the Pinery Limestone composing the lower part of the Bell Canyon Formation at the top of Nipple Hill.

One noteworthy fact is that the suggested boundaries are poorly characterized by other fossil groups. The procedure for their establishment seems to have been restricted to "the driving of conodont spikes" without taking into consideration the evolution of biota as a whole. Therefore, most of the boundaries appeared to lie within the traditional stages. For example, as mentioned above, the Cathedralian–Hessian boundary was lowered to the base of the Hessian Stage (as Ross interpreted its scope) (Ross and Ross, 1987). The conodont-based Roadian–Wordian boundary occurs within the Roadian Stage, as is traditionally accepted, and the lower Capitanian boundary is placed much higher than the first appearance of *Polydiexodina* and, probably, *Timorites*, although it was previously substantiated by the appearance of these genera. As a rule, the suggested boundaries are also discordant with the stratigraphic levels of events in the development of the basin, some of which may be of subglobal character. For example, a significant unconformity at the base of the Vidrio Beds is likely to correspond to the unconformity at the base

of the Midian Stage in many Tethyan sections. Previously, these beds were considered to be the base of the Capitan Formation (Capitanian Stage) (Ross, 1986; Ross and Ross, 1995), but now they are included in the Wordian Stage and the unconformity appears to occur inside the latter. Everything said above indicates the artificial nature of the suggested boundaries (except for the lower and upper boundaries of the Guadalupe Group), which are hardly traceable outside the Delaware basin. This is why we have no doubts concerning the establishment of the Guadalupe Group, but mistrust the stage boundaries.

As mentioned above, the International Subcommission on Permian Stratigraphy met during the symposium. The meeting was attended by eight SCPS members and all participants of the symposium. The meeting was to discuss the three-member division of the Permian System as it was accepted in Washington (table). There was no discussion of those stages of the Operational Scheme suggested to be constituent units of the series. The exceptions were the Kungurian and Dzhulfian stages, whose introduction into the scale raised doubts.

The division of the Permian System into three series (Cisuralian, Guadalupian, and Lopingian) with their boundaries corresponding to basal levels of the conodont *J. nankingensis* and *C. postbitteri–C. crofti* zones did not elicit any discussion, and the majority of the votes proposed it for ratification by the International Commission on Stratigraphy.

There were debates on the Kungurian Stage. Glenister argued for the Cathedralian Stage and its stratotype in the Glass Mountains to be accepted in its place. His main argument was that the North American sections of this stage had an advantage over the Uralian section as they were entirely composed of marine facies and showed distinct conodont-based boundaries. According to another opinion, there are no reasons for substituting the Kungurian Stage with the Cathedralian, because both stages have equal scopes. It would be logical to recognize all stages of the Cisuralian Series in the Uralian stratotype region, the stages of the Guadalupian Series in the Delaware basin, and stages of the Lopingian Series in southern China. If none of the Uralian sections demonstrate a distinct lower Kungurian boundary, the boundary stratotype can be chosen in the North American sections. The participants voted for introducing this stage into the scale under the double name *"Cathedralian-Kungurian"* and selecting its stratotype in the Glass Mountain section.

The Dzhulfian Stage was also a subject of discussion. Jin Yugan sharply objected to recognizing any stages of the Lopingian Series outside southern China. The voting majority supported his opinion and the Dzhulfian Stage was not approved.

The decisions on these problems were voted on not only by the SCPS members but by all those present. Therefore, the results of the vote cannot be considered to be official decisions of the SCPS. Therefore, a second meeting of the SCPS was held to put these problems to a vote among SCPS members only. The second round of voting gave a majority to the Kungurian Stage and rejected the Dzhulfian.

As a result, the variant of the Operational Scheme without the Dzhulfian Stage has been approved (table). After the symposium it was distributed to the SCPS members for postal ballot. The results of the voting were announced at the SCPS meeting during the 30th Session of the International Geological Congress in Beijing, August 1996. The division of the Permian System into three series has been unanimously approved. As for the stages, they were preliminarily accepted. However, a final decision is impossible untill their boundaries and the boundary-stratotypes are correctly defined. For the next four years, efforts of the SCPS will be concentrated on these problems and the accepted stage scale may be changed.

In conclusion, we would like to say a few words as to how to assess the decisions taken in Alpine. It is obvious that they do not satisfy most Russian specialists on the Permian who are against any attempts to revise the traditional Uralian scale. This was confirmed by the stratigraphic workshop held at the Geological Institute of the Russian Academy of Sciences to discuss the results of the Second Guadalupian Symposium. The reaction of the opponents of the accepted variant of the scale was extremely negative and sometimes very emotional. However, no positive proposals have been made beyond the statement on "the great stratigraphic potential of the Uralian sections." But we may ask why this potential has not been realized although the Uralian scale has existed for many decades. The answer is well known: specific features of the Uralian sections, which are mostly represented by continental facies and lack distinct biostratigraphic boundaries between stages, prevent tracing these boundaries outside the stratotype region. One cannot but admit that the Upper Permian stages of the Uralian scale are not adjustable to the standard Permian scale, because neither these stages nor the boundaries between them meet the international requirements concerning stratigraphic units of this rank and their boundaries. We think that efforts should be directed toward retaining all Uralian stages in the lower series, particularly the Kungurian Stage, instead of reanimating the entire Uralian scale. An important but more remote task is to subdivide the Lower Permian into two independent series according to the natural-historical basis of the scale. For this purpose, it was suggested to organize a special working group of the SCPS and to carry out additional paleontological studies of the Artinskian Stage.

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Reviewer A.S. Alekseev

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Permian–Triassic Boundary Deposits in Framing Structures of the East Siberian Platform

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Abstract—The first description of the stratotype of the Ust'-Kel'ter Horizon with its specific characteristics is given. A new stratigraphic unit, the Marininskii Horizon, is distinguished and substantiated, and its stratotype is described. The Marininskii and Ust'-Kel'ter horizons continue upward the succession of continental Permian–Triassic stratigraphic units of Siberia. Their recognition allows us to address the problem of age and correlation of the Permian–Triassic boundary deposits in the northern Siberian platform and its framing structures.

Key words: *continental deposits, Marininskii Horizon, Ust'-Kel'ter Horizon, Permian, Triassic, Siberian platform*

In the last few years, it was established that the Siberian platform and its framing structures comprise a much more complete succession of Permian–Triassic boundary deposits than found in Europe. The lower part of this succession, which consists of basic tuffs, tuffaceous–sedimentary rocks and basalts, is paleontologically best-characterized in the Tunguska syneclise. Here, the section comprises the Tutonchana, Dvorogii, and Putorana horizons (*Resheniya* ..., 1991). The Dvurogii Horizon was later divided into the Lebedev and Khungtukun horizons (Sadovnikov and Orlova, 1995). The Marininskii and Ust'-Kel'ter horizons supplement the regional stratigraphic scheme of continental deposits in Middle Siberia adopted by the Third Interdepartmental Meeting (table).

The Ust'-Kel'ter Horizon, which we singled out in 1981, comprises the lower, mostly terrigenous part of the Triassic continental sequence of the northern Siberian

Correlation of regional stratigraphic horizons of the Permian–Triassic deposits in Russia

System	Series	Stage	East European platform	Siberian platform and its framing*		Eastern slope of the Western Verkhoyansk Range
Triassic	lower	Olenekian	Vokhma**			*Hedenstroemia hedenstroemi*
		Induan			Ust'-Kel'ter	*Vavilovites* spp.
						Glyptophiceras nielseni
					?	*Otoceras boreale*
?	?	?			Marininskii	?
Permian	upper	Taimyrian***		Putorana		
		Tatarian	Vyatka	Dvurogii****	Khungtukun	
					Lebedev	
				Tutonchana		
			Severnaya Dvina	Gagarii Ostrov		
				Degali	Ust'-Degali	

Notes: * According to resolution of the Third Interdepartmental Stratigraphic Meeting, 1979 (left column), and proposed in this paper (right column).

** Excluding the Astashikha Formation, which probably corresponds to part of the Putorana Horizon.

*** Our proposal of 1994.

**** At the Third Interdepartmental Stratigraphic Meeting, the subdivision of the Dvurogii Horizon into two subhorizons proposed by G.N.Sadovnikov was supported by participants, but was not accepted by the Interdepartmental Stratigraphic Commission.

platform and its framing (Sadovnikov *et al.*, 1981). The horizon was named after the Ust'-Kel'ter Formation whose stratotype is located in the lower course of the Kel'ter River, the left tributary of the Nuora River, which flows into the Tumara River (right tributary of the Aldan River in its lower course). At that time, the stratotype of the horizon was poorly studied and paleontologically poorly substantiated. The studies in the stratotype area in the Kel'ter River lower course (Sadovnikov and Orlova, 1995) led us to the conclusion that one more unit (the Marininskii Horizon) should be singled out between the Ust'-Kel'ter Horizon and gray-colored deposits known to be Permian in age (Sadovnikov and Orlova, 1993). This paper presents geological materials that were collected in the areas near the mouth of the Kel'ter River, the adjacent Nuora River, and its lower left tributary, the Mol River. These materials significantly supplement the paleontologic characteristics of the Ust'-Kel'ter Horizon and, along with data from the Taimyr sequence (Sadovnikov and Orlova, 1990), justify the validity of the Marininskii Horizon. Below, we give the first description of stratotypes of the Marininskii and Ust'-Kel'ter horizons. This section is of great importance because it is one of the most complete sections of nonmarine Permian–Triassic boundary deposits known anywhere. Therefore, it is precisely here, rather than in incomplete European sections, that the boundary between the Permian and Triassic systems can be defined in nonmarine facies, and continental analogues of the Dorshamian Stage can be established (Fig. 1).

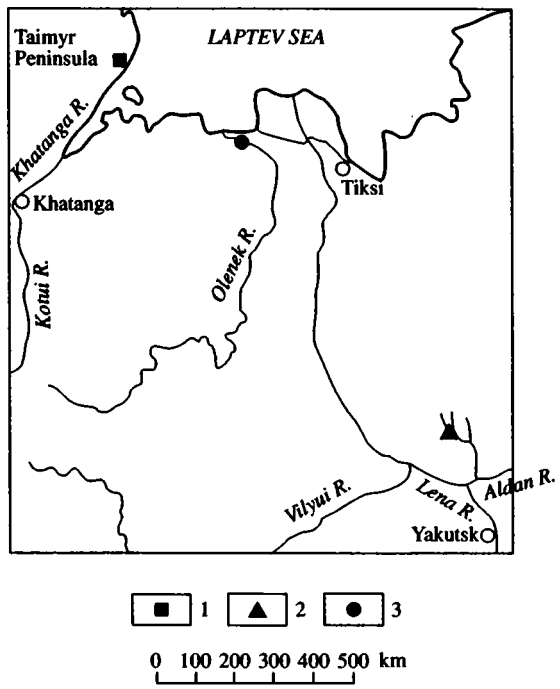

Fig. 1. Location of stratotype sections. 1, the Marininskii Horizon; 2, the Ust'-Kel'ter Horizon; 3, the Olenekian Stage.

Overlying terrigenous deposits are distributed along the Siberian continent periphery: in the Taimyr and western Verkhoyansk regions. Central Siberia is a region where two units of the Lower Triassic, the under-Olenekian Beds, later named the Induan Stage, and the Olenekian Stage, were distinguished for the first time. The region hosts stratotypes of regional ammonite zones of the Induan Stage unambiguously correlated with those of the marine standard of the stage. Thus, this region is promising for correlation between stratigraphic scales of marine and nonmarine Permian–Triassic boundary deposits.

Determinations of fossils were performed by E.F. Orlova (conchostracan), G.N. Sadovnikov (plant remains), A.S. Dagys (ammonites), and N.K. Kurushin (bivalves).

The oldest rocks exposed in the Kel'ter River lower course are the Permian deposits represented by gray and black siltstones, brownish-gray and light-colored, fine- and, less commonly, medium-grained, polymictic sandstones displaying cross, cross-lenticular, and, rarely, parallel bedding, and enclosing subordinate layers of black argillites with remains of small densely-veined *Rufloria* leaves.

Up the section, a 70-m-thick member of gray fine- to medium-grained sandstones and gray, greenish-gray, and greenish-black siltstones occurs. The lower barren part of the member (about 15 m) is mostly composed of sandstones and is best exposed in the left side of the Kel'ter River 6.9 km up from the mouth of the large nameless right tributary. Grains in sandstones are represented by quartz, feldspar (20–40% each), and basic to acidic volcanics (15–30% each). The heavy fraction is dominated by turmaline (25–30%). Also common are garnet (10–30%), epidote (0–28%), Cr-spinel (0–27%) and ore minerals (1–29%); subordinate minerals are zircon (10–14%), apatite, and rutile (up to 10% each). The matrix is of hydromica–chlorite and, partly, of secondary carbonate composition.

The middle part of the member (about 15 m) consists predominantly of greenish-gray siltstones with shale and rare sandstone interlayers. Siltstones in the right bank of the Kel'ter River, 1.2 km upstream of its mouth, contain remains of conchostracan species *Falsisca* ex gr. *turaica* (Nov.), *F.* aff. *verchojanica* (Mol.), *Lioestheria ignatjevi* Nov., *L. blomi* Nov., *Sphaerestheria aldanensis* Nov., and *Cyclestheria rossica*. Siltstones exposed in the west-southwestern slope of the mountain (bench mark 1647), 0.8 km below its summit (the left side of the Nuora River downstream of the Erekit River), yielded plant remains: *Tomiostrobus bulbosus* Sadovn., *T.* cf. *gorskyi* (Vlad.) Sadovn., *T. pachyphyllus* Sadovn., *Cladophlebis* sp., and *Carpolithes* sp. Plant fragments of *Tomiostrobus* cf. *belozerovii* Sadovn., *Neocalamites* sp., *Radicites* sp., and *Cladophlebis* sp. were sampled in similar rocks outcropping

in the left bank of the Kel'ter River, 2.7 km up from its mouth.

The upper part of the member (about 40 m) is mostly composed of gray and greenish-gray siltstones with shale and fine-grained sandstone interlayers; in the lower part of the member, ellipsoidal and oblate concretions and lenses of silty limestones are observed. These rocks are best exposed in the right bank of the Kel'ter River, 6.3 km upstream of a large right tributary. Here, limestones bear conchostracan remains: *Wetlugites pronus* Nov., *Sphaerestheria* cf. *aldanensis* Nov., cf. *Cyclotunguzites gutta* (Lutk.), *Eulimnadia* ex gr. *bergessensis* Nov., *E.* cf. *mutusjaensis* Mol., *Limnadia* ex gr. *komiana* Mol., *Falsisca* sp., and *Glyptoasmussia quadrata* Nov. Similar siltstones in the left bank of the Kel'ter River, 2.4 km from its mouth, incorporate plant remains cf. *Tomiostrobus* sp. and conchostracan: *Falsisca* sp. nov., *Cornia sileenica* Mol., *C. vosini* Mol., *Pseudestheria sibirica* Nov., and *Cyclestheria rossica* Nov. In the right bank of the Kel'ter River, 2 km from its mouth, the upper part of the siltstone member contains remains of *Tomiostrobus pachyphyllus* Sadovn.

Plant and conchostracan fossils allow the gray-colored unit to be correlated with the Marininskii Formation of eastern Taimyr (Sadovnikov and Orlova, 1990; Sadovnikov *et al.*, 1981) and with the Nedzhelin Formation of the western slope of the West Verkhoyansk Range (Sadovnikov *et al.*,1981). We suggested to single out these deposits as the independent Marininskii Horizon (Sadovnikov and Orlova, 1993) accepting as its stratotype that of the synonymous formation exposed along the Marininskii Creek in the eastern Taimyr Peninsula. According to V.P. Belozerov, the following rock succession is exposed along this small creek (mainly in its right bank) from the base upsection:

(1) Fine rhythmical alternation of black to dark-gray siltstones, gray fine-grained sandstones, gray sandy siltstones, and black shales with plant remains of *Pleuromeia* cf. *taimyrica* Sadovn., *Tomiostrobus bulbosus* Sadovn., *T. deltoideus* Sadovn., *T. latus* Sadovn., *Neokoretrophyllites* sp., *Todites* sp., *Lepidopteris arctica* Mog., *Peltaspermum* sp., *Caprolithes* ex gr. *minor* Pryn., and cf. *Podozamites*(?) sp. The thickness is 120 m.

(2) Dark-gray, splintery shales interlayering with thin beds of fine-grained sandstones, which bear plant remains: *Tomiostrobus* cf. *bulbosus* Sadovn., *Neokoretrophyllites*(?) sp., *Lepidopteris* cf. *arctica* Mog., *Carpolithes* sp., and cf. *Phiridopsis* sp. The thickness is 40 m.

(3) Fine rhythmic alternation of dark-gray siltstones, brownish in the upper part, with fine-grained sandstones and shales. The thickness is 12 m.

(4) Fine rhythmic alternation of dark-, lilac-, and greenish-gray siltstones and greenish- to pinkish-gray siltstones and shales. Three thick (2.5, 4, and 1 m) beds of greenish- and pinkish-gray sandstones occur respectively at the base, in the upper part and at the top of the member. They yielded plant remains: *Pleuromeia* cf. *taimyrica* Sadovn., *Tomiostrobus belozerovii* Sadovn., *T. bulbosus* Sadovn., *T. fusiformis* Sadovn., *T.* cf. *gorskyi* (Vlad.), *T. latus* Sadovn., *Neocalamites*(?) sp., *Cladophlebis* sp., *Rhizomopteris* sp., *Lepidopteris arctica* Mog., and *Peltaspermum* sp. as well as conchostracan *Falsisca* sp. The thickness is 26 m.

In addition to the Marininskii Formation of Taimyr, the Marininskii Horizon also comprises the Sobolokh and Yulegir units and the Nedzhelin Formation of the western slope of the West Verkhoyansk Ridge. The horizon bears *Cornia sileenica* Mol., *Falsisca verchojanica* Mol., *Tomiostrobus belozerovii* Sadovn., *T. bulbosus* Sadovn., and *T. deltoideus* Sadovn.

In the Kel'ter River lower course, the described gray-colored member is overlain by variegated terrigenous deposits best exposed in the left bank of the Kel'ter River, 2.4 km downstream of its mouth (Fig. 2). Here, in the middle part of the outcrop, the bed occurrence is almost vertical and somewhat undulating; in the outcrop marginal part, the dip angle is up to 40° (Fig. 3). The cyclicity patterns and paleontologic data show that rocks here form a syncline with the oldest overturned beds outcropping in its marginal parts. The most complete section is observed in the western limb of the syncline. Here, the following rock succession is exposed (from the base upward):

(1) Light-gray, medium-grained sandstones grading higher into gray and greenish-gray fine-grained sandstone. The thickness is 15 m.

The member 1 and underlying member 2 are divided by an unexposed interval 7 m wide.

(2) Siltstones, greenish-gray, coarse-grained, banded, micaceous, with interlayers of clayey siltstones and fine-grained platy sandstones. The thickness is 6.8 m.

The members 1 and 2 most likely belong to the Marininskii Horizon.

(3) Cherry-red and green sandy siltstones interlayering with shales. The thickness is 3.4 m.

(4) Fine-grained sandstones interlayering with variegated sandy, sometimes clayey, laminated, thin-platy, micaceous siltstones. The thickness is 1.7 m.

(5) Variegated, banded, clayey and sandy siltstones with a sandstone bed (5 cm) near the top and interlayers and lenses of shales (up to 10 cm). The thickness is 8.7 m.

Siltstones occurring 2.7 m higher from the base of the last member bear conchostracan remains: *Falsisca* ex gr. *turaica* (Nov.), *F.* ex gr. *verchojanica* Mol., *Cornia sileenica* Mol. *C. vosini* Mol., *Sphaerestheria aldanensis* Nov., *Wetlugites pronus* Nov., *Pseudestheria sibirica* Nov., single *Limnadia* sp., *Lioestheria* cf. *blomi* Nov., *L. ignatjevi* Nov., *Concherisma* sp., *Loxomicroglypta* cf. *balbukensis* Nov., and *Glyptoasmussia* sp. Upsection (5.6 m above the base), plant remains of cf. *Tomiostrobus*(?) sp. occur.

(6) Light-green splintery shales. The thickness is 0.9 m.

(7) Sandstones, greenish-gray and cherry-red, fine-grained, cross-bedded, banded, massive, in places platy. The thickness is 11.2 m.

(8) Variegated, fine-grained, banded, massive sandstones with conchoastracan remains: *Wetlugites pronus* Nov., *Lioestheria blomi* Nov., rare *Cyclestheria rossica* Nov., *Sphaerestheria aldanensis* Nov., *Pseudestheria sibirica* Nov., *Falsisca* aff. *turaica* (Nov.), single *Cyclotunguzites gutta* (Lutk.), *Pseudestheria* aff. *tumariana* Nov., *Glyptoasmussia* sp., and *Loxomicroglypta* sp. The thickness is 1 m.

(9) Sandy or clayey, sometimes micaceous siltstones, which are massive or platy, and splintery. The thickness is 4 m.

(10) Dark-gray sandy shales. The thickness is 1.6 m.

(11) Clayey micaceous siltstones, gray, platy, with splintery jointing. The thickness is 1 m.

(12) Alternation of greenish-gray to cherry-red, banded, platy, often micaceous, fine-grained sandstones and sandy siltstones. In the eastern limb of the fold, these rocks contain plant remains of *Tomiostrobus latus* Sadovn. and *Carpolithes*(?) sp. The thickness is 12.4 m.

(13) Greenish-gray, massive sandy and platy clayey siltstones, in places grading into fine-grained sandstones. The thickness is 2.6 m.

(14) Siltstones, greenish-gray and cherry-red, banded, clayey and sandy, often micaceous. The thickness is 5 m.

In the eastern limb of the fold, the thickness of siltstones corresponding to members 13 and 14 increases up to 31.3 m.

(15) Greenish-gray, laminated and thin-platy, sandy siltstones with imprints of *Tomiostrobus* cf. *migayi* (Schved.) and conchostracan remains: *Wetlugites pronus* Nov., *Cyclestheria rossica* Nov., *Cyclotunguzites gutta* (Lutk.), *Sphaerestheria* sp., *Pseudestheria sibirica* Nov., *P.* aff. *tumariana* Nov., *Lioestheria* cf. *ignatjevi* Nov., and *Concherisma* sp. The thickness is 0.35 m.

(16) Greenish-gray, banded siltstones, massive at the base and top of the member and platy in its middle part. Middle siltstones yielded poorly preserved plant remains and fragments of conchostracan *Lioestheria ignatjevi* Nov., *L. blomi* Nov., *Pseudestheria* cf. *sibirica* Nov., *Wetlugites pronus* Nov., and *Glyptoasmussia quadrata* Nov. The thickness is 1.4 m.

(17) Cherry-red, banded, platy, micaceous siltstones with interlayers and lenses (5–25 cm thick) of clayey siltstones. In the middle part of the member, a calcareous siltstone bed (7 cm) occurs. The thickness is 2.2 m.

(18) Greenish-gray, banded, platy, fine-grained siltstone with a bed (3 cm) of clayey siltstones. The thickness is 0.6 m.

(19) Dark-gray clayey siltstones. The thickness is 0.1 m.

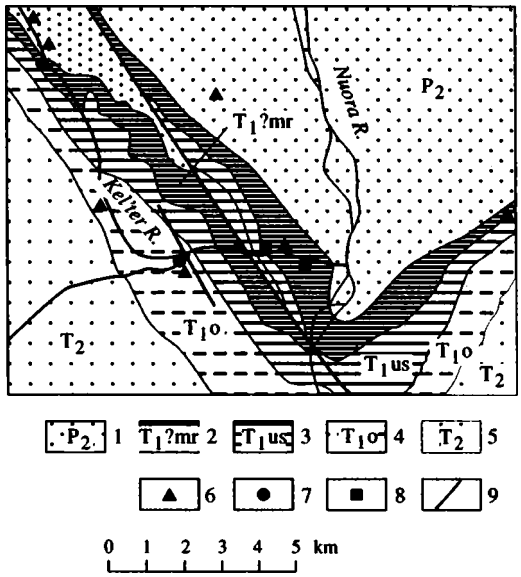

Fig. 2. Geological map of the Kel'ter River lower course: (1) Upper Permian deposits; (2) the Marininskii Horizon; (3) the Ust'-Kel'ter Horizon; (4) the Olenekian Stage; (5) Middle Triassic deposits; (6) finds of conchostracan fossils; (7) finds of plant remains; (8) stratotype of the Ust'-Kel'ter Horizon; (9) faults.

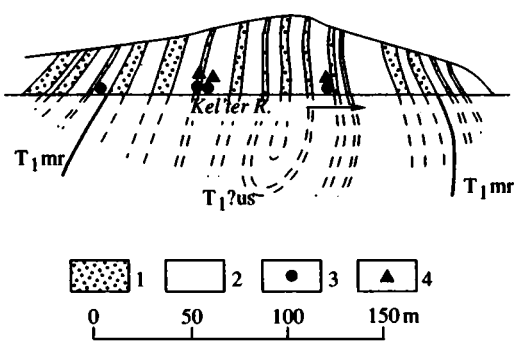

Fig. 3. Stratotype section of the Ust'-Kel'ter Horizon: (1) sandstones and sandy siltstones; (2) clayey siltstones and shales; (3) finds of conchostracan fossils; (4) finds of plant remains.

(20) Greenish-gray, sometimes cherry-red, banded, platy, clayey siltstones with interbeds (5–30 cm) of sandy siltstones. The upper layers of the member incorporate poorly preserved remains of poorly preserved conchostracan *Lioestheria* sp., *Glyptoasmussia* sp., and plants *Tomiostrobus*(?) sp. The thickness is 4.2 m.

(21) Greenish-gray and cherry-red, banded, massive, sandy siltstones. The thickness is 0.9 m.

(22) Greenish-gray, banded, thin-platy, splintery, micaceous siltstones with interlayers and lenses of clayey siltstones. Rocks are ferruginous in the upper part of the member. The thickness is about 5 m.

(23) Conglomerates composed of large (1–3 cm across) flattened pebbles of shales and siltstones with a silty matrix. The thickness is 0.1 m.

(24) Sandy siltstones, greenish-gray, banded, massive or platy. The thickness is 5.3 m.

(25) Clayey to sandy micaceous siltstones, gray, greenish-gray to cherry-red, banded, platy, with splintery jointing. In similar rocks exposed in the eastern limb of the fold, the diverse conchostracan assemblage includes: *Lioestheria blomi* Nov., *L.* aff. *blomi* Nov., *L. ignatjevi* Nov., *Pseudestheria sibirica* Nov., *Cyclestheria rossica* Nov., *Sphaerestheria aldanensis* Nov., *S.* aff. *aldanensis* Nov., *Cyclotunguzites gutta* (Lutk.), *Wetlugites pronus* Nov., *Glyptoasmussia quadrata* Nov., *Loxomicroglypta subcircularis* (Chern.), *Concherisma* sp., *Falsisca* sp. (aff. *turaica* Nov.), *Limnadia* sp., *Eulimnadia* cf. *sotiana* Nov., *E.* cf. *mutusjaensis* Mol., *Cornia* cf. *sileenica* Mol., and *Estherina* sp. The thickness is 2.1 m.

(26) Sandy siltstones, greenish-gray and cherry-red, banded, massive. The thickness is 0.2–0.3 m.

(27) Sandy siltstones, cherry-red (at the base) and greenish- gray (at the top), laminated, locally cross-bedded. The thickness is 0.2 m.

(28) Shales, cherry-red (at the base) and greenish-gray (at the top), splintery. The thickness is 0.2 m.

(29) Cherry-red and greenish-gray, platy, sandy micaceous siltstones, which become black and clayey at the top of the member (upper 3–4 cm). The thickness is 1.7 m.

(30) Grenish-gray, banded, massive, sandy siltstones with a bed (2–3 cm) containing flat shale pebbles and remains of plants *Tomiostrobus* cf. *migai* (Schved.). The thickness is 0.25 cm.

(31) Cherry-red, platy siltstones. The thickness is 1.2 m.

(32) Gray, dark-gray, and greenish-gray, platy siltstones. The thickness is 1.5 m.

In the eastern limb of the fold, the total thickness of siltstones corresponding to members 31 and 32 is 7.1 m.

(33) Greenish-gray and cherry-red, locally micaceous siltstones. Rocks are horizontally bedded, massive or platy, with ripple marks at the lower surface. In several layers up to 15 cm thick (two layers in the lower part of the member and three layers in its upper part), they are sandy, loamy, and platy. The heavy fraction is dominated by turmaline (30%), garnet (24%), and Cr-spinel (19%); apatite (10%) and zircon (9%) occur in subordinate amounts. Also present are rutile (6%) and epidote. The thickness is 6.2 m.

Higher layers of the section outcrop in the left bank of the Kel'ter River, 0.6 km downstream of the mouth of a large right tributary after a gap in exposures that cannot be reliably estimated.

(34) Gray and greenish-gray fine- and medium-grained sandstones crowned with a 0.4-m-thick bed of gritstones bearing shale pebbles. Rocks are composed of quartz (30%), feldspar (30%), acidic (20%) and basic (10%) volcanics, quartzite, chert (5%), and biotite (5%), whose grains are cemented by a filmy and interstitial silica–hydromica–chlorite matrix. The heavy fraction is mainly represented by epidote (38%), Cr-spinel (20%), and garnet (18%); subordinate minerals are zircon (12%) and apatite (10%). Turmaline and rutile occur as accessory minerals. The thickness is 6.1 m.

(35) Greenish-gray and pinkish, platy, clayey siltstones. The thickness is 3.7 m.

(36) Greenish-gray, clayey and sandy siltstones. The thickness is 6.2 m.

(37) Greenish-gray shales, turning upsection cherry-red and, further, pinkish-gray. The upper part of the member incorporates a bed of sandy siltstone (0.3 m). The heavy fraction from the latter is dominated by garnet (32%) and zircon (30%), subordinate are turmaline (16%) and rutile (10%) while apatite (6%), Cr-spinel (3%), and epidote (2%) are accessory minerals. The thickness is 4 m.

(38) Fine- and medium-grained sandstones, gray, platy in the lower part and pinkish, massive in the upper part of the member. The uppermost 0.5-m-thick bed contains abundant and large shale pebbles. The thickness is 4.3 m.

(39) Greenish-gray and variegated (in the upper part), sandy siltstones interlayering with dark-gray, banded shales. The thickness is 3.6 m.

(40) Gray, fine- to medium-grained, massive sandstones resting upon the uneven surface of member 6. The thickness is 0.8–0.9 m. Sandstones corresponding, evidently, to members 5–7 are exposed in the right bank of the Kel'ter River, directly downstream from the second right tributary, where their thickness reaches up to 12 m.

(41) Dark-gray, clayey siltstones. The heavy fraction is dominated by epidote (20%), turmaline (20%), zircon (19%); less common is apatite (10%), Cr-spinel (7%), and rutile (5%) are also present. The thickness is 5 m.

(42) Variegated, fine- to medium-grained sandstones with a dark-gray, clayey siltstone interlayer (0.2–0.4 m) located 1.2 m above the base of the member. Mineral grains are represented by quartz (30%), feldspar (30%), basic (30%) to acidic (10%) volcanics, microquartzite, and biotite embeded into a ferruginous chlorite matrix. The heavy fraction is mostly composed of abundant turmaline (83%), subordinate epidote (10%), rare zircon (4%), and accessory ore minerals, apatite, rutile, and garnet. The thickness is 11.1–12 m.

(43) Greenish-gray, clayey siltstones with shale interbeds. The thickness is 8–10 m.

(44) Cherry-red and pinkish, fine-grained sandstones. The thickness is 15 m. Most likely, sandstones corresponding to this member are also exposed in the right bank of the Kel'ter River, 0.2 km downstream of the

large right tributary where, after an insignificant gap in outcrops, they are overlain by the following succession:

(45) Gray, clayey and sandy siltstones with calcareous concretions and interlayers of gray, fine-grained sandstones and gray shales. The thickness is 10 m.

After a gap in outcrops downstream of the right tributary mouth, the exposed beds are represented by:

(46) Gray and greenish-gray, fine-grained sandstones. The thickness is 3 m.

(47) Gray, clayey and sandy siltstones interlayering with black shales and gray, medium-grained sandstones. The heavy fraction is completely represented by epidote. The thickness is about 10 m. Similar rocks in the left bank of the Kel'ter River across the next (downstream) right valley bear plant remains of *Tomistrobus* cf. *migayi* (Schved.), *Neocalamites* sp., and *Radicites* sp.

The upper layers of the Ust'-Kel'ter Horizon are exposed in the left bank of the Kel'ter River, 3.5 km upstream of the mouth of the large right tributary.

(48) Variegated, mainly cherry-red shales and siltstones with ellipsoidal carbonate concretions. The thickness is 15 m.

(49) Greenish-gray, clayey siltstones with spheroidal carbonate concretions. The thickness is 30 m.

(50) Variegated, mainly cherry-red, clayey siltstones. The thickness is 11 m.

(51) Cherry-red (lower 10 m) and greenish-gray (upper 12 m) siltstones with interlayers of greenish-gray and cherry-red shales. In the upper part of the member, siltstones become gray and clayey. The thickness is 25 m.

We propose to consider the Ust'-Kel'ter Horizon as a part of the above-described variegated unit below the gray-colored Marininskii deposits, although earlier we also included them into this horizon (Sadovnikov *et al.*, 1981). In this case, the Vostochnyi Taimyr Formation and its analogues should also be excluded from the Ust'-Kel'ter Horizon.

With the proposed scope, the Ust'-Kel'ter Horizon comprises the Fad'yukuda, Keshina, and Tsvetkova Mys formations of the Taimyr region and Sagynkan Formation of the western Verkhoyansk region. The horizon is marked by such characteristic fossils as *Wetlugites pronus* Nov., *Lioestheria ignatjevi* Nov., *Pseudestheria sibirica* Nov., *Cyclestheria rossica* Nov., *Cyclotunguzites gutta* (Lutk.), *Tomiostrobus latus* Sadovn., and *T. migayi* (Schved.).

In the lower course of the Kel'ter River, the above-described variegated deposits are overlain by black, though sometimes brownish or reddish shales with abundant concretions, locally confined to individual horizons, and with lenses of silty limestones. The rocks are best exposed in the right bank of the river near the mouth of its large right tributary. They yielded remains of conchostracan *Glyptoasmussia quadrata* Nov. and *Wetlugites pronus* Nov. that already belong to the Olenekian Stage.

The Ust'-Kel'ter conchostracan assemblage widely distributed in the eastern slope of the Verkhoyansk Range and in the Delin'ya River valley associates with *Wordieoceras* forms, and thus corresponds to the upper parts of lower Induan. The Marininskii conchostracan assemblage is commonly considered to be Induan in age. This is quite possible though not beyond doubt. Deposits with the Marininskii conchostracan and floral assemblages, widespread in the western slope of the Verkhoyansk Range, are not recorded in the eastern slope. Molin (1965) has found a similar conchostracan assemblage in the only locality of the Dulgalakh River valley near the mouth of the Magan-Khaya River. It is not inconceivable that the rare occurrence of the assemblage in this area is related to erosion of the uppermost Marininskii deposits. No *Otoceras* forms are yet found together with determinable conchostracan, therefore the possibility that these deposits are younger than the Marininskii Horizon cannot be ruled out (Sadovnikov and Orlova, 1993).

It is interesting to mention here the Syurbelyakh Formation, which was singled out in the northern Verkhoyansk Range (Vavilov *et al.*, 1981) and is recently thought to be of Induan age because it incorporates a floral assemblage of the Korvunchan type (Biterman *et al.*, 1977). Earlier, we showed that the Syurbelyakh Formation comprises heterochronous deposits (Sadovnikov *et al.*, 1981). In the Unguokhtakh River valley, flora of the Kornunchan type was found in the lowermost part of these deposits, which are gray-colored and resemble those of the Yulegir sequence underlying the Nedzhelin Formation in the south. In turn, they are overlain, most likely with a hiatus, by variegated deposits bearing the Marininskii conchostracan assemblage.

Consequently, the Syurbelyakh Formation represents a combined unit and is thus not valid in the proposed stratigraphic range. Therefore, it is improper to correlate the Ungoukhtakh flora with the Korvunchan assemblage in order to substantiate its Induan age.

Thus, in the area near the Kel'ter River mouth, two stratigraphic units distinctly differing in lithological and paleontological characteristics can be distinguished between gray-colored deposits known to be of Permian age and black shales of the Olenekian Stage. According to paleontological data, these units, which are well traceable over the eastern slope of the Verkhoyansk Range and in the Taimyr Peninsula region, can be treated as independent stratigraphic units: the Marininskii and Ust'-Kel'ter horizons.

Reviewer Yu.B. Gladenkov

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The Decision on the Precambrian–Cambrian Boundary Stratotype: A Breakthrough or Misleading Action?

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"When the only tool you own is a hammer, every problem begins to resemble a nail"

Abraham Maslow

Abstract—The official decision to chose a section in the Burin Peninsula, Newfoundland, as the global stratotype for the lower Cambrian boundary and to define this boundary by the appearance of the trace fossils *Trichophycus* (=*Phycodes*) *pedum* was made without evaluating the real biostratigraphic potential of ichnofossils. Their potential is far less than that of body skeletal fossils used for subdividing the entire Phanerozoic. Correct interregional correlations of the Burin model of the lower Cambrian boundary cannot be provided by the available biostratigraphic and isotopic-geochronologic records, and lithological characters of the selected type section exclude the application of C- and Sr-isotopic chemostratigraphic criteria for this purpose. These facts have actually repudiated the decision taken on the Global Stratotype Section and Point for this boundary. The choice made is of no use for real stratigraphic practice.

Key words: *Precambrian, Proterozoic, Vendian, Cambrian, Nemakit-Daldynian Stage, Tommotian Stage, boundary, boundary stratotype, ichnofossils, small shelly fossils, acritarchs, chemostratigraphy, isotopic age*

At the 29th session of the International Geological Congress (IGC) in Kyoto, 1992, the International Union of Geological Sciences (IUGS) ratified the proposal of the International Commission on Stratigraphy on Global Stratotype Section and Point (GSSP) for the Precambrian–Cambrian boundary or, to be exact, for the lower boundary of the Cambrian System (Brasier *et al.*, 1994; Landing, 1994). The section in the Burin Peninsula, southeastern Newfoundland, was chosen as the stratotype, and the stratigraphic point "was selected with the guidance from the level marked by the lowest occurrence of *Phycodes pedum* (a trace fossil), at the base of the Cambrian *Phycodes pedum* Biozone" (Brasier *et al.*, 1994, p. 6).

Now, when the euphoria of the formal accomplishment of 25 years of work is over (Cowie, 1978, 1985, 1992; Cowie and Rozanov, 1983; Brasier *et al.*, 1994; Debrenne and Debrenne, 1995), it has become obvious that this decision is erroneous, if not fatal. This concerns the chosen basis for the GSSP definition—both the Burin section proper and ichnofossils (but not skeletal fossils), which are notorious for their low potential for correlation.

I

The lower Cambrian boundary has attracted the attention of researchers for over 40 years. For this period great progress has been made from the belief in the impossibility to define a level of the boundary and to chose even a possible stratotype region (*Colloque ...*, 1958) to the understanding that three regions only (the Siberian platform, Yangtze platform, and Avalon zone of Newfoundland) are suitable for this purpose. Moreover, at a certain moment there were minor differences of opinion on the criteria for drawing this boundary and its level. At the meeting of the Working Group on the IGCP Project 29 "The Precambrian–Cambrian Boundary" held in Paris, 1974, it was agreed that the Ediacara-type fauna should be considered as Precambrian, and the "olenellid/fallotaspid" trilobite faunas should be considered as Cambrian (Cowie and Rozanov, 1983). A few years later, at the meeting in Cambridge, 1978, the Working Group agreed that the boundary stratotype should be placed as close as practicable to the base of a stratigraphic unit with rich Tommotian fosill fauna (*s.l.*) (Cowie, 1978).

Later the three main candidates for the global boundary stratotype: Ulakhan-Sulugur (the Aldan River in Siberia), Meishucun (Yunnan Province, China), and the Burin Peninsula (Newfoundland, Canada) were proposed. For some time, the members of the Working Group were engaged in comparing the strong and weak points of these possible stratotypes. In 1983 the decision seemed to have been reached: the interested specialists who met in Bristol (UK) concluded that the Ulakhan-Sulugur section best met the international standard. However, the subsequent postal ballot did not give this section a majority of the votes necessary for its ratification. The following drawbacks of the section were reported: the lack of a reliable object suitable for isotopic dating, an abrubt facial change and an inconformity close to the proposed boundary point, the unclear genesis of Bed 8 (see below) at whose base the "golden spike" was supposed to be driven, and inaccessibility. The Newfoundland section was the second in the ballot. However, in 1984 the Chinese section alone was put to a postal ballot of the Working Group and was approved by nearly 80% of the votes (Cowie, 1985).

At the 27th Session of the IGC in Moscow, 1984, the Chinese section was proposed to the ICS for discussion as a global Precambrian–Cambrian boundary standard. However, the commission suggested to postpone the decision, taking into account the ambiguous biostratigraphic and chemostratigraphic correlations of the proposed level (Point B of the Meishucun section), insufficiently accurate paleontological and isotopic-geochronological substantiation, and possible hiatuses evident from lithological and paleontological data. The IUGS supported this suggestion of the International Commission on Stratigraphy (Cowie, 1965). At the 28th session of the IGC in Washington, 1989, the Working Group confirmed the objections against the Meishucun section as the boundary stratotype. It was decided to discuss this problem again at the 3rd International Symposium on the Cambrian System.

During this symposium held in Novosibirsk, 1990, the stratigraphy of the Precambrian–Cambrian boundary deposits was discussed in several reports and at a special meeting of the Working Group. However, a careful consideration of the problem did not take place. The decision was actually predetermined and the Working Group chairman insisted on voting on the Newfoundland section to complete the procedure.

The Working Group examined the Burin section two times (in 1978 and 1987), but untill 1983 it was not taken seriously as a candidate for the global Precambrian–Cambrian boundary standard. After the first examination, the Working Group members stated that despite the finely exposed distinct succession of strata, the Burin section was not adequately studied and, like the majority of siliciclastic sections, was paleontologically poorly characterized. Due to the consequent energetic efforts of T. Fletcher, G. Narbonne, E. Landing and their colleagues, it became one of the well studied sections (Narbonne *et al.*, 1987; Landing 1992, 1994; and references therein). Later the Burin section was visited by more than 30 specialists; most of them were representatives of North America. No weak points of the section have been officially documented. However, opinions of the participants of the excursion were clearly expressed in the questionaire (Cowie and Brasier, 1989, p. 207). Thirteen specialists estimated the potential of the section for global correlation as being moderate or poor; only eight persons confirmed the occurrence of well preserved and diverse fauna, whereas 22 specialists were skeptical about the paleontological characteristics of the section. Moreover, 12 participants doubted the very idea of the GSSP choice for the lower Cambrian boundary in predominantly siliciclastic sequence characterized by trace fossils of uncertain systematic affinity.

Despite the skepticism, the Burin section remained the main candidate for the global standard of the Precambrian–Cambrian boundary. The reason was that many members of the Working Group reestimated the stratigraphic potential of small shelly fossils (SSF), which have long been considered as promising much for standartization of the lower Cambrian boundary (Rozanov *et al.*, 1969; Cowie, 1978, 1985; *The Precambrian–Cambrian ...*, 1989; Rozanov, 1984; *Yarusnoe ...*, 1984). By the late 1980s, an opinion has been formed that this fauna is characterized by strong provincialism. Being restricted to shallow-water carbonate platforms, the SSF cannot be used for subdivision and correlation of the Vendian–Cambrian siliciclastic deposits, which have broader global distribution and yield specific paleontological remains.

At that time some active members of the Working Group (Narbonne *et al.*, 1987; Narbonne and Myrow, 1988; Landing *et al.*, 1989; and others) suggested to use ichnofossils for the boundary definition. The suggestion was based on the identification of some successive ichnozones in the Precambrian–Cambrian boundary deposits (Crimes, 1987). Two ichnozones were of particular interest for substantiating the boundary level: the younger *Trichophycus pedum* (=*Phycodes pedum*) Zone that contained the index-species together with some other typical Cambrian ichnofossils and the underlying *Harlaniella podolica* Zone characterized by less diverse and, what is more important, simpler ichnofossils. These were the reasons why some members of the Working Group admitted a possibility to define the lower Cambrian boundary by a change of these zones, i.e., by the appearance of trace producers with Paleozoic aspect. Unfortunately, adherents of this opinion did not take into account that more than half of the reference sections in different continents, which were enlisted by Cowie (1992), were carbonate-type successions that were paleontologically better studied than siliciclastic sections. Sufficiently diverse trace-fossil assemblages were found in 20 siliciclastic sections (Narbonne *et al.*, 1987; Crimes, 1992), four of which

were missing in Cowie's list. This is one third less than the number of sections bearing the SSF in this interval.

In the winter of 1990–1991, the postal ballot resulted in approval of the Burin section as the global standard of the Precambrian–Cambrian boundary; 52% of voting members of the Working Group were in favor, but all the Russian and Chinese specialists voted against, and 60% of voting members of the IUGS also supported this decision.

II

The GSSP choice in the Burin Peninsula draws our attention to three points of prime importance.

First, no boundary of the global Phanerozoic chronostratigraphic scale (the Cambrian base is the lower boundary of the Phanerozoic) has ever been defined by ichnofossils. Every boundary is substantiated by remains of rapidly evolving skeletal fauna. Moreover, trace fossils are not as widely used in biostratigraphy as body fossils. By their nature, ichnofossils represent the results of interaction between an organism and environments, which are imprinted in sediments and reflect the behavior and locomotion of the animal rather than its morphology. That is why no trace producer, with rare exception, can be related to a particular taxonomically defined animal. It is well known that different organisms can produce identical traces and, *vice versa,* several forms of ichnofossils may be traces of a single organism (Fedonkin, 1987; Crimes, 1987). Partly for this reason, the trace fossil taxonomy is still incomplete and does not have a multiorder hierarchy similar to that of zoological taxonomy. Being a reflection of animal behavior, a trace fossil consists of many elements, each reflecting a certain state of an organism (Fedonkin, 1981; Jensen, 1993). The occurrences of all these elements in identical microfacies imply a "conspecificity" of ichnofossils and their possible usage as a correct but auxiliary biostratigraphic tool. Obviously, this could be possible in the future. Biostratigraphic disadvantages of paleoichnology also include the evident facies-controlled distribution of ichnofossils, which reflects the paleoecology of producers and the taphonomy of their traces. Striking differences in diversity and composition of ichnocoenoses from carbonate and siliciclastic successions (the first are usually rich in skeletal fossils and very poor in ichnofossils) also do not count in favor of traces. All these features indicate a very moderate (if not poor) biostratigraphic potential of trace fossils. Due to the very nature of ichnofossils, it is impossible to identify their natural-historical and evolutionary successions that can serve as a basis for biostratigraphic schemes.

Second, the fact should be taken into account that ichnofossils exhibit far less potential for the stratigraphic resolution of the Vendian–Cambrian boundary interval than SSF. For example, two zones of skeletal fossils correspond to a lower part of the single *Rusophycus avalonensis* Ichnozone. It is unlikely that the recognition of an independent ichnozone between the *Trichophycus pedum* and *Rusophycus avalonensis* zones will provide a more refined and reliable biostratigraphic resolution of these deposits than that based on the zonal succession of SSF *Anabarites trisulcatus–Purella antiqua* (*"Purella"*)*–Nochoroicyathus sunnaginicus* (Khomentovskii and Karlova, 1991, 1992, 1994a). It was because of the high-resolution of this scale that the universal isochroneity of the boundary between the SSF *trisulcatus* and *antiqua* zones was doubted (Khomentovskii and Karlova, 1994a). Later, the doubts were supported by C-isotopic chemostratigraphic data (Kaufman *et al.*, 1996). Obviously, such a high resolution ability is far beyond the real potential of "ichnofossil" biostratigraphy. Another noteworthy fact is that genera and species of ichnofossils are commonly one order less in number than those of skeletal fossils known from the coeval successions.

Third, the definition of the lower Cambrian boundary through the appearance of trace fossils inevitably gives rise to an untrivial problem of correlating scales and boundaries defined by ichnofossils and those based on skeletal fossils (small shelly fossils, archaeothyans, trilobites). The problem is of prime importance because Cambrian biostratigraphy is entirely based on skeletal fossils. A particular but challenging aspect of this problem is a time-interrelation between the lower Cambrian boundary substantiated by ichnofossills in the Burian section and the lower boundaries of the Nemakit-Daldynian and Tommotian stages of Siberia, which are defined by skeletal fossils. As we shall see below, there is still no definite solution of this problem and, hence, the application of the Burin model of the boundary to the carbonate sections remains unproven; the model immanentely lacks the potential to be universal.

The attempts to use ichnofossils for defining the global stratotype point of the Precambrian–Cambrian boundary were warranted by references to their widest distribution in siliciclastic sections and subglobal succession of zonal assemblages, in particular those of *Harlaniella podolica* and *Trichophycus* (=*Phycodes*) *pedum.* Landing wrote: "Trace fossils are present to the near exclusion of shelly fossils in the siliciclastic-dominated sequences on most continents in the boundary interval (i.e., Avalon, Australia, Baltica-East European platform, Hercynian Europe, Laurentia, Gondwana), and the ichnofossils in the Precambrian–Cambrian boundary interval ... are practical for interregional correlation." (Landing, 1994, p. 180).

However, this statement contradicts the thesis of Narbonne *et al.*, (1987) about the strong provincialism of the Cambrian faunas. If the Early Cambrian organisms were displaying provincialism, how could their traces be cosmopolitan? Even if we admit that some traces were of a global facies-unrestricted distribution, they could not be produced by the same organisms.

The statement regarding the high potential of ichnofossils for the correlation of the boundary interval is not in accordance with the fact that the succession of *Harlaniella podolica–Trichophycus pedum* zonal assemblages has been recorded in only two regions (Podoliya of the Ukraine and Newfoundland). The Burin section only shows the immediate succession of these two forms, whereas in Podoliya their stratigraphic ranges are separated by the 17-m-thick Okunets Formation lacking these ichnofossils (Gureev, 1988). An impoverished paleontological assemblage from the Podoliya deposits referrable to the *Harlaniella podolica* Zone should also be mentioned. There are sporadic Ediacara-type fossils and scarce, taxonomically uniform ichnofossils (Fedonkin, 1987; Gureev, 1988; *The Vendian ...*, 1990). This probably indicates a general freshening of the Kotlin paleobasin of the East European platform and some shallowing of its Ukrainian part (Korenchuk *et al.*, 1993; Pirrus, 1992). Even if we leave the impoverishment aside, the fact remains that we can observe the same succession of ichnozones and correlate the same level in two sections only, one of which is a stratotype. The oldest Cambrian trace-fossil assemblages with *Trichophycus pedum* from all other sections are variable both in composition and age. In the doubtless situations, for example, in northern Spain and southern Australia, these assemblages are dated as the late Atdabanian and late Tommotian–early Botomian respectively (Crimes and Anderson, 1985; Zhuravlev and Gravestock, 1994). However, in the Kashmir the earliest *Phycodes* sp. occurs together with *Redlichia*, but the earliest *Ph. pedum* is recorded upsection together with *Ptychoparia* (De, 1993). Therefore, the definition of a certain level separating doubtless Vendian and Cambrian ichnofossil assemblages in regional successions is an independent problem, whose solution may not be correct. The difficulties are related to the usually incomplete regional ichnofossil successions and the restricted biotopes of their producers.

Particular attention should be given to the Vendian-Cambrian boundary deposits of the Olenek uplift, which is actually the only region, where rich SSF assemblages of the Nemakit-Daldynian and Tommotian stages are associated with diverse ichnofossils (Fedonkin, 1985; *The Vendian ...*, 1990; Khomentovskii and Karlova, 1992). The lower, pre-Tommotian part of the Kessyusa Formation contains abundant trace fossils predominantly represented, however, either by new taxa or by forms known from both the *podolica* and *pedum* zones. At the same level there is *Didymaulichnus tirasiensis* Palij described from the Rovno Horizon (Regiostage) of Podoliya (Palij, 1976) and *Phycodes* sp. (Fedonkin, 1985, Plate 23, Fig. 5), but the identification of the latter form as the ichnospecies *Ph. pedum* seems doubtful. The characteristic Cambrian *Rhizocorallium* occurs 20 m below the appearance of the typical Tommotian SSF assemblages in the Kessyusa Formation. However, the significance of this form for the correlation of the Siberian sections with the chosen GSSP is negligible because it is absent from the Burin section. In the Olenek uplift, the Nemakit-Daldynian trace-fossil assemblage including *Phycodes, Gordia, Olenichnus, Planispiralichnus, Protospiralichnus,* and *Dydimaulichnus tirasiensis* Fed. is succeeded by large meandering forms *Didymaulichnus meanderiformis* Fed. and *Diplocraterion* sp. of the upper, Tommotian part of the Kessyusa Formation. This ichnofossil succession has not been established in the Burin section. This implies the incompleteness of either the section or its paleontological characteristics. In the Meishucun section of China the ichnofossil taxa characteristic of Zone III of Crimes (1987) appear at Point "B," which marks the base of the Tommotian equivalents (Xing *et al.*, 1991), while the earliest *Trichophycus* (=*Phycodes*) *pedum* was only found 8 m downsection.

Thus, the occurrence of the distinct laterally persistent discrete assemblage of the *Trichophycus pedum* Zone, which is recognizable in the upsection increasing diversity of the Vendian-Cambrian ichnofossils, needs to be further documented. According to the available data, the immediate replacement of *Harlaniella podolica* by *Trichophycus pedum*, which is observed in the Burin section, seems to be a unique phenomenon. In the North Siberian and South Chinese sections, the early Tommotian SSF appear so close to the appearance of the Cambrian ichnofossils that delimitation of a stratigraphic unit bounded by these levels is evidently beyond the resolving power of ichnofossil biostratigraphy.

III

The identification of such a unit in the Burin section (Narbonne *et al.*, 1987; Landing, 1992, 1994; Brasier *et al.*, 1994) rests on the assumptions that there is a significant pre-Tommotian hiatus over the most part of the Siberian platform (including the Tommotian stratotype region) and that the lower boundaries of the Tommotian Stage and the *pedum* Zone in the Newfoundland succession are separated by a noticeable interval (see below) extending from the middle part of the Cuslett Formation down to the lower part of Member 2 of the Chapel Island Formation (Narbonne *et al.*, 1987; Landing, 1994).

The first assumption disagrees with the continuous faunal evolution in the Nemakit-Daldynian and Tommotian times, which is evident from the early (Rozanov *et al.*, 1969) and later (Val'kov, 1987; Missarzhevskii, 1989) studies on the corresponding faunas, regardless of the different views of the authors on the correlation of the Precambrian–Cambrian boundary beds in the northern and southeastern parts of the Siberian platform. It has recently been shown that in the Uchur-Maya region, SSF significantly increase in abundance and diversity in the pre-Tommotian deposits of the extensive boundary area located between the transitional facial zone that includes the Upper Vendian and Lower Cambrian sections in the Aldan River area and

the eastern facial zone embracing the Uchur and Maya river basins (Khomentovskii and Karlova, 1991, 1992, 1993, 1994a). Two successive *Anabarites trisulcatus* and *Purella antiqua* (*"Purella"*) zones were recognized there below the lower Tommotian *Nochoroicyathus sunnaginicus* Zone. As for the eastern facial zone (the Olenek uplift and Anabar massif), most of the present authors believe that the characteristic forms of the *sunnaginicus* Zone appear there throughout the entire Tommotian Stage. By the way, this is one of the reasons for the diverging opinions on the biostratigraphic correlation of the Vendian–Cambrian boundary deposits in the northern and southeastern parts of the Siberian platform (Savitskii, 1975; Rozanov, 1982, Val'kov, 1987l; Missarzhevskii, 1989; Repina and Rozanov, 1992; Khomentovskii and Karlova, 1992, 1994a).

However, the fact remains that in the Tommotian stratotype region (the Aldan River area) and in most of the Uchur-Maya region, there is a distinct unconformity at the base of the Pestrotsvet Formation, which is evidenced by the fact that this formation rests on different horizons of the underlying Ust'-Yudoma Formation (Semikhatov and Serebryakov, 1983; Starnikov *et al.*, 1983; Sukhorukov, 1984; Khomentovskii and Karlova, 1992, 1993, 1994b). The significance of this unconformity increases in view of the mentioned ambiguous interpretation of the origin of Bed 8 of the Ulakhan-Sulugur section (assumed to be a sedimentary layer *in situ*, or karst fillings in the pre-Pestrotsvet relief), which, by definition, defines the position of the lower Tommotian boundary (Rozanov *et al.*, 1969). According to the C-isotopic chemostratigraphic records in the Vendian–Cambrian deposits of the Aldan and Anabar successions, some researchers, including M.A. Semikhatiov, demonstrated a significant duration of the pre-Pestrotsvet (pre-Tommotian) hiatus (Knoll *et al.*, 1995; Kaufman *et al.*, 1996). However, in Russian publications the hiatus is commonly taken to be of a rather small range, judging from the successive character of the Tommotian and Nemakit-Daldynian faunas, and, if *in situ* position of Bed 8 is accepted, then also of successive composition of faunas from the basal Pestrotsvet Formation and this bed. In any case, in many Siberian sections (the Anabar massif, Olenek and Sukharikha uplifts) the most probable level for the lower Tommotian boundary does not coincide with the break in sedimentation.

As for the opinion of Narbonne, Landing, Brasier and their co-authors concerning the relative positions of the lower boundaries of the *pedum* Zone and the Tommotian Stage in the Burin section (Narbonne *et al.*, 1987; Brasier *et al.*, 1994; Landing, 1994), it stems from a disputable interpretation of very poor SSF assemblages of this section and, as we shall see below, can be questioned.

In the Avalon section, reliable stratigraphic markers are beds with trilobites *Serrodiscus, Ladadiscus* and *Calodiscus* (Landing, 1992) in the middle part of the Brigus Formation correlated by means of this fauna with the lower Botomian Stage. The older beds of this section, separated by an unconformity from the Brigus Formation, constitute four conformable formations (in discending order): Foster Point, Cuslett, West Center Cove, and Petley. The first formation and the uppermost part of the second belong to the regional *Camenella baltica* Zone. This zone can be logically referred to the Atdabanian Stage on the basis of the occurrence of: (a) zooproblematics *Rhombocorniculum cancellatum* and hyolite *Gracilitheca*, which appear first in the Atdabanian deposits, and (b) brachiopode *Micromitra* unknown from the Tommotian deposits. The co-occurrence of the forms passing from the older deposits cannot serve as evidence for assigning this zone to the pre-Atdabanian level. The conclusion of Landing and co-authors concerning the pre-Tommotian age of the underlying *Sunnaginia imbricata* Zone, which comprises the lower Cuslett Formation (Member 1d and downward) and the West Center Cove and Petley formations, is not well substantiated as well because *Sunnaginia imbricata* and *Torellella laevigata* found there are not recorded below the Tommotian base and are very common in the upper part of the Tommotian successions in the eastern Siberian platform (*Yarusnoe* ..., 1984; Khomentovskii and Karlova, 1992).

In the Burin section, the older SSF assemblage is separated from that of the *Sunnaginia imbricata* Zone by the Petley Formation deposited above the pre-Petley unconformity and the Random Formation lacking skeletal fossils. In the opinion of most of the present authors, this assemblage, which is supposed to define the *Watsonella crosbyi* Zone and is characteristic of two upper members of the underlying Chapel Island Formation (Narbonne *et al.*, 1987; Landing, 1992, 1994), cannot be regarded as pre-Tommotian as well. This opinion is supported by the following facts: (1) *Watsonella* (=*Heraultipegma*) does not occur below the Tommotian base in southeastern Siberia and is found in the lower Tommotian beds in northern Siberia; (2) *Allatheca degeeri* and *Lapwortella* from the *Watsonella crosbyi* Zone of Newfoundland appear in the middle and lower Tommotian deposits of Siberia respectively (*Kembrii* ..., 1992; Khomentovskii and Karlova, 1992, 1994a); (3) the Wrekin Quartzites of Shropsire, which are considered to be equivalents of the Randon Formation in the British part of the Avalon platform (Landing, in press), contains the Dominopole (Lyukati) acritarch assemblage (Wright *et al.*, 1993). Thus, both the Wrekin Quartzites and Random Formation cannot be older than the upper Tommotian (even in the interpretation of Vidal and Moczydlowska, see Vidal *et al.*, 1995) and most likely belong to the lower Atdabanian. It should be emphasized that the references to the occurrence of the Lyukati acritarch assemblage in the *regularis* Zone of the Khalaulakh section of northern Siberia (Rudavskaya and Vasil'eva, 1984; Vidal *et al.*, 1995) must be interpreted with an account of the presence of *Pagetiellus* sp. at the "Tommotian" level of this section

(Vidal *et al.*, 1995). This form proves the enclosing deposits to be no older than the third zone of the Atdabanian Stage. Even the lowest *"Ladatheca" cylindrica* Zone of the Burin section established by SSF does not seem to be of pre-Tommotian age. In Siberia *"Ladatheca"* recorded from this zone (if its identification is correct) appears in the uppermost horizons of the deposits commonly regarded as pre-Tommotian in age, and is abundant in the Tommotian beds.

Taking into account that the SSF assemblages of the Burin section are extremely impoverished and have a discrete distribution in the section, we can take the suggested correlations with the Siberian section as provisional. Nevertheless, they seem to be more substantiated and more likely than the speculative correlations of Landing and his colleagues, who draw the lower Tommotian boundary in the Burin section at the base of the *Camenella baltica* Zone, which we refer to the Atdabanian Stage. The above does not mean that we question the difficulty in the interregional correlation of the Vendian-Cambrian boundary beds by means of SSF. This difficulty is comparable with those met in substantiating and correlating the Phanerozoic stratigraphic boundaries by many other Phanerozoic fossil groups. The mentioned fauna includes particularly difficult groups whose biostratigraphic potential has yet to be defined. However, this fauna also includes some index forms (e.g., *Lapwortella tortuosa*) and the forms (e.g., *Lapwortella bella, Rhombocorniculum insolutum, R. cancellatum, Mobergella radiolata,* etc.) that serve as regional and interregional markers.

One could believe that a telecorrelation of the Burin GSSP is possible by means of acritarchs, the more so in that they provide the subdivision and correlation of the classical Vendian–Cambrian seccessions of the Russian platform (e.g., Volkova, 1985; Moczydlowska, 1991). Alas, the repeated attempts to discover acritarchs in the Vendian–Cambrian deposits of Newfoundland were not successful. However, even if some representative acritarch assemblages were found there, they could not serve as a tool for the correlation of the Vendian–Cambrian boundary beds between the sections other than siliciclastic, because the acritarch-based correlation of these beds between the Russian and Siberian platforms is the subject of much controversy.

IV

It is evident from the above that there are no reliable biostratigraphic criteria to correlate not only the Precambrian–Cambrian boundary defined in the Burin section, but also the entire pre-Atdabanian part of that section. Nonbiostratigraphic methods are unable to solve the problem as well.

The almost exclusively siliciclastic composition of the Vendian–Cambrian boundary beds in the Burin section and their relatively strong secondary alterations prevent employing Sr- and C-isotopic chemostratigraphy for correlation of the chosen boundary. The analysis of the O- and C-isotopic compositions in the carbonate nodules and carbonate interbeds occurring in the pre-Atdabanian part of the Burin section shows that the original isotopic systems of these rocks were reset by both the thermal effect during deep burial and the influence of post-Cambrian granitic intrusions, and by the vital effect, that is the bacterial decay of organic carbon in anaerobic environments (Brasier *et al.*, 1992). Therefore, as Brasier and co-authors emphasized, this section cannot even offer a prospect for chemostratigraphic correlations based on the isotopic compositions of carbonate and organic carbon (Brasier *et al.*, 1992, p. 270). Perspectives for the application of Sr-isotopic chemostratigraphy to the Precambrian–Cambrian boundary interval in Newfoundland are also equal to zero due to both the mentioned thermal effect and the mode of carbonate occurrence—nodules and thin beds within the siliciclastic succession. Such occurrence inevitably causes an intense Sr-isotopic exchange between carbonate and siliciclastic fractions that alters the original isotopic signatures of carbonates (Vinogradov *et al.*, 1994; Gorokhov *et al.*, 1995). Studies on sequence stratigraphy also suggest the low potential of the Burin section for correlation because of the inconsiderable lateral extension of sequence boundaries (Myrow and Hiscott, 1993).

In principle, the recent precision of the U–Pb dating of volcanogenic zircons opens a possibility for solving the correlation problems by means of isotopic-geochronological data. However, the scarcity of the relevant datings for the Vendian–Cambrian boundary beds and the uncertain stratigraphic position of some dated rocks prevent the application of this method to the sections under consideration. The following U–Pb zircon ages are most significant in the context of this article.

(1) Zircons from pebbles of the conglomerates occurring near the base of the Tyuser Formation in the Kharaulakh uplift, northern Siberia, have the weighted mean $^{207}Pb/^{206}Pb$ age of 534.6 ± 0.4 Ma (Bowring *et al.*, 1993). The conglomerates are conformably overlain by carbonates with a rich SSF assemblage, which is referred either to the lower Tommotian zone (Missarzhevskii, 1989; Bowring *et al.*, 1993) or, more frequently, to the mid-Tommotian *Dokidocyathus regularis* Zone (e.g., Repina *et al.*, 1994; *Kembrii ...*, 1992). Thus, this value defines the maximal age limit of this zone. Within the analytial error, it is consistent with the refined U–Pb zircon age (530 ± 5 Ma) from the Baiyanshao Member of the Meishucun type section in China (Sambridge and Compston, 1994). This member, which immediately predates the level of occurrence of the diverse SSF assemblage (Point B), belongs to the upper part of the *Anabarites–Protoherzina* Zone and may be regarded as pre-Tommotian or, in any case, pre-middle Tommotian in age. The isotopic age of the upper Tommotian horizons can be inferred from the zircon date (521 ± 7 Ma) obtained for the upper Tommotian–lower

Atdabanian succession of Morocco (Compston *et al.*, 1992).

(2) Zircons from the volcanic breccia near the base of the Kessyusa Formation in the Olenek uplift, northern Siberia, have the weighted mean $^{207}Pb/^{206}Pb$ age of 543.9 ± 0.24 Ma (Bowring *et al.*, 1993). The SSF confined to the overlying parts of the Kessyusa Formation define their assignment to the Nemakit-Daldynian *Anabarites trisulcatus* and *Purella antiqua* zones and the Tommotian *Nochoroicyatus sunnaginicus* Zone (Khomentovskii and Karlova, 1992, 1993). The biostratigraphic (*op. cit.*) and chemostratigraphic (Kaufman *et al.*, 1996) data show that the lower Kessyusa deposits are correlative with the lower horizon of the Nemakit-Daldynian stratotype in the western slope of the Anabar massif. Thus, following Bowring *et al.* (1993), the date of 543.9 ± 0.2 Ma can be considered as determing the isotopic age of the lower Nemakit-Daldynian boundary. However, Khomentovskii and Yakshin questioned the volcanogenic origin of the Kessyusa breccia because they believe the breccia to be similar in composition to rocks related to diatremes occurring in pre- and post-Kessyusa deposits. However, whatever the genesis of the Kessyusa breccia may be, its U–Pb age cannot be used for the isotopic dating of the lower Cambrian boundary defined in the Burin section, as done by Bowring and co-authors. The elucidation of chronological and stratigraphic relationships between this boundary and the lower boundary of the Nemakit-Daldynian Stage is a specific problem, which has not yet been finally solved (see above).

(3) Zircons from the tuff bed correlative with the Chapel Island Formation near the town of Saint Johns, New Brunswick, have the upper intersect U–Pb age of 531.0 + 1.8/–1.1 Ma and the weighted mean $^{207}Pb/^{206}Pb$ age of 530.7 ± 0.9 Ma. The latter value is considered to be the best estimation of the time of the tuff deposition (Isachsen *et al.*, 1994). Interregional correlations and paleontological data show that the tuff bed belongs to the middle part of the *Rusophycus avalonensis* Ichnozone or, to be exact, to the SSF *Watsonella crosbyi* Zone (Landing, 1994; Isachsen *et al.*, 1994; Brasier *et al.*, 1994), which is stratigraphically higher than the boundary between the *Trichophycus* (=*Phycodes*) *pelum* and *Harlaniella podolica* ichnozones of the Burin section. Thus, the age of the lower Cambrian boundary accepted in that section exceeds 530.7 ± 0.9 Ma.

Contrary to the opinion of Isachsen *et al.*, (1994), this $^{207}Pb/^{206}Pb$ zircon age does not mean that the estimated isotopic age of the lower Tommotian boundary by Bowring *et al.*, (1993) is too old and should be revised. Isachsen and co-authors proceeded from the assumption that both the *Watsonella crosbyi* Zone and the significant overlying part of the Burin section are pre-Tommotian in age (Landing, 1991, 1994; Brasier *et al.*, 1994). As we showed above, this assumption can be doubted and the above zone is likely to be of Tommotian age. Thus, the date of 530.7 ± 0.9 Ma cannot be used for the clarification of the temporal relationships between the lower boundaries of the Nemakit-Daldynian Stage and the *Trichophycus pedum* Zone. If the inferred Tommotian age for the *Watsonella crosbyi* Zone is correct, the age of the lower Tommotian boundary appears to be older than the mentioned value. It should be noted that the isochron Rb–Sr age of a fine (0.1–0.2 μm) fraction of authigenic illite from the upper part of the Lontova Horizon (Regiostage) of Estonia is 538 ± 8 Ma (Gorokhov *et al.*, 1994).

(4) According to SHRIMP II ion microprobe data, zircons from the tuffs of the Mistaken Point Formation, southeastern Newfoundland, which is stratigraphically below the horizons under discussion and bears Ediacara-type fossils, fall into three equal groups corresponding to 607 ± 1; 591 ± 5; and 571 ± 4 Ma (Compston, 1994). In Compston's opinion, the last date is the best estimation of the age of the tuffs. Previously, the U–Pb zircon age of this formation, which was obtained by the traditional method, was reported to be 563 ± 3 Ma (Benus, 1988). However, analytical data required to support this date were never published. So, the isotopic-geochronological data indicate that the Burin GSSP has a maximal age limit of 560–570 Ma.

(5) Zircons from the Slowatycze tuffs of eastern Poland have the $^{207}Pb/^{206}Pb$ age of 551 ± 4 Ma (Compston *et al.*, 1995). The tuffs occur near the top of the Volyn' Group of the Vendian age and grade upsection into the siliciclastic Semyatyche and Beloe Pole formations overlain by beds correlative with the Kotlin Horizon (Regiostage) (Moczydlowska, 1991). Some researchers (e.g., Velikanov, 1988) include the Beloe Pole–Semyatyche succession into this horizon. In other words, stratigraphically the Slowatyche tuffs are located well below the level that corresponds to the Kotlin-Rovno boundary and is marked by the replacement of *Trichophycus pedum* by *Harlaniella podolica* in the Dniester section. Thus, if this replacement happened geologically simultaneously both in the Dniester and Burin sections, the above date allows us to refine the estimated maximal age limits of the Burin model for the Precambrian–Cambrian boundary and to consider it to be 551 ± 1 Ma.

Thus, the available isotopic-geochronological data do not contribute to the understanding of the time relations between the lower boundaries of the Nemakit-Daldynian and Tommotian stages, on the one hand, and the Burin GSSP for the lower Cambrian boundary, on the other hand. In our opinion, this data shows only that (a) the lower Cambrian boundary accepted in the Burin section is younger than 551 ± 4 and older than 530.7 ± 0.9 Ma; (b) the lower boundary of the *Dokidocyathus regularis* Zone is younger than 534.6 ± 0.5 and older than 521 ± 7 Ma; (c) this boundary is older than 530.7 ± 0.9 Ma, if the suggested stratigraphic level of the *Watsonella crosbyi* Zone of Newfoundland is correct; (d) the age of the lower Nemakit-Daldynian boundary is close to 543.9 ± 0.24 Ma. This inference

based on the U–Pb zircon ages is in disagreement with the recent Sm–Nd isochron dating of phosphate shells and collophane from the upper part of the *Anabarites–Protoherzina* Zone of China (Yang *et al.*, 1996). The last age values, which appeared after our paper had been submitted for publication, are 562.8 ± 7.9, 562.1 ± 5.7, and 570.3 ± 17.1 Ma. The first two values are based on different numbers of samples from a single section—the Meishucun stratotype (Beds 3–7 of the Yuhucun Formation), while the third value was calculated by using samples from the same beds and their analogues in the Sichuan Province and northwestern China. The discussion of the correctness of the reported Sm–Nd ages is beyond the scope of this publication. However, it should be noted that they are inconsistent with the U–Pb zircon ages of the upper horizons of this zone in the Meishucun section (see above).

V

Thus, the available biostratigraphic and isotopic-geochronologic data does not provide reliable telecorrelations of the Burin GSSP for the Precambrian–Cambrian boundary, and the lithologic features of the Precambrian–Cambrian boundary beds of southeastern Newfoundland exclude the possibility to use C- and Sr-isotopic chemostratigraphic criteria for substantiation of these correlations. This evident conclusion in effect repudiates the decision to select the GSSP for the lower Cambrian boundary in the Burin Peninsula section, Newfoundland.

In this situation, it would be most expedient to appreciate all the realities related to the selected GSSP for the lower Cambrian boundary and to begin again special international studies on the comparative evaluation and correlation of regional stratotypes of the Vendian–Cambrian boundary beds. The objects to be studied must include carbonate successions of the Uchur-Maya region, the Olenek uplift and Anabar massif, and the successions of siliciclastic type in the western Russian platform, which is known for its acritarch zonation. Likely, the Meishucun section is not the only one in China that is worth further attention. This seems to be the only possibility for a successful choice of GSSP for the lower Cambrian boundary in the future.

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Middle Jurassic Basins of the Eastern Mediterranean and Their Microbiota (Foraminifers and Ostracodes)

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Abstract—The effect of different abiotic factors on the evolution and distribution of bottom microbiota, namely, foraminifers and ostracodes, in the Middle Jurassic basins of the Eastern Mediterranean is discussed. The research is based on data on Jurassic sediments in Syria, where more than 90 sections of that age were studied in different regions. The relationship between foraminifer and ostracode distribution and bathymetric zones of the basin was elucidated. The studied groups of benthic organisms were demonstrated to be most responsive to a change in the character and granulometric composition of sediments, and also to hydrodynamics of bottom water masses, sedimentation rate, water salinity, and calcium carbonate saturation. All these parameters are in turn directly connected to the depth of a certain part of the basin and its distance from the shoreline. The greatest taxonomic diversity of benthic foraminifer assemblages was recorded in the sublittoral zone, especially in its outer and middle parts, whereas ostracodes were most diverse in the middle and inner sublittoral zone. Fluctuations of water salinity in the studied basins were also dependent on the distance from the shoreline and a bathymetric factor. In nearshore shoals, the salinity varied to a greater extent as indicated by the distribution of stenohaline forms in ostracode assemblages. A distinct bathymetric differentiation was recorded in the taxonomic composition of foraminifers, which were represented almost entirely by cosmopolitan forms in deeper-water parts of the basin, namely, in epibathyal, outer and middle sublittoral zones, whereas Tethyan endemics clearly dominated in inner sublittoral and littoral shoals.

Key words: *foraminifers, ostracodes, Middle Jurassic, abiotic factors, bathymetric zones, epibathyal, sublittoral, littoral, taxonomic diversity*

The study of sedimentation in marine basins includes both stratigraphic research of certain intervals of a sequence and the elucidation of environmental conditions for the biota that depicted all changes in ancient basins. Paleoenvironmental conditions had an important bearing on the taxonomic composition of paleobiocoenoses of certain groups, namely, foraminifers and ostracodes. They also affected the relationship between faunal components, morphology of shells, structure of assemblages, and were responsible for the predominance of a certain reproduction type. According to ecologic characteristics, the studied Jurassic foraminifers and ostracodes are attributed to the benthic epifauna, which inhabited upper layers or the surface of sediments. In other words, they inhabited the oxidizing zone, the thickness of which varied slightly in relatively stable conditions of the epicontinental sea. All environmental changes had a direct impact on the rearrangement of faunal assemblages, which were used to distinguish stages in the development of Jurassic foraminifers associated with geologic events in the Eastern Mediterranean (Kuznetsova, 1994; Kuznetsova and Dobrova, 1995). It turns out that rearrangements of assemblages directly dependent on abiotic factors were substantially more frequent than evolutionary changes caused by other factors, such as biological peculiarities of faunas, their morphotypes, their rate of development, intensity of speciation, tolerance degree, and others.

Biological evolution of faunas is the subject of a separate research project, whereas the purpose of this paper is to analyze the connection and dependence of foraminifer and ostracode assemblages on abiotic factors in Jurassic basins.

Shelf basins of the Eastern Mediterranean are extremely interesting subjects for such investigation. Their sediments were inhabited by abundant biota formed under specific conditions of warm shoals and water masses highly saturated with calcium carbonate.

Analyzing abiotic factors most significant for the sedimentation and benthic fauna development, we should distinguish those which can be reliably studied by using fossils. Among these factors are (1) the tectonic regime responsible for the geologic structure of a sedimentation basin; (2) the mobility of bottom waters affecting structure and granulometric composition of sediments; (3) the productivity and saturation of water with phytoplankton, which is responsible for the trophic factor; (4) water temperature inferable from indicative forms of phytoplankton, namely, coccolithophores and warm-water organisms such as corals and algal bioherms; (5) the chemical composition and

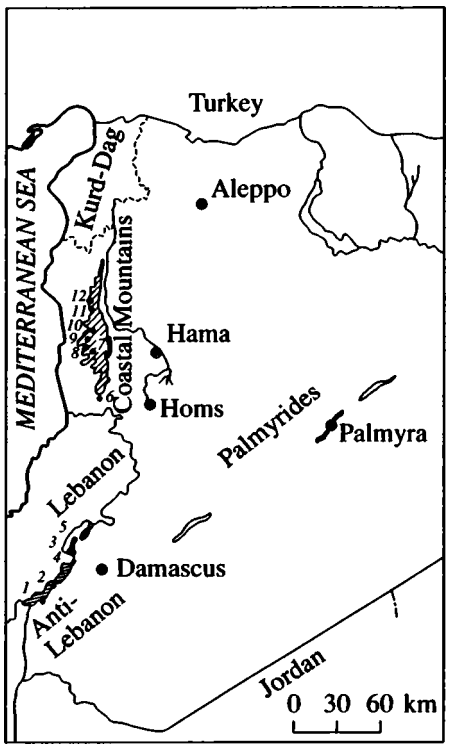

Fig. 1. Location of the studied sections of Jurassic sediments in Syria. Anti-Lebanon: (*1*) Hadar; (*2*) Arne; (*3*) Wadi Al Karn; (*4*) Sad Al Karn; (*5*) Rowda. Coastal Mountains: (*6*) Nicola Quarry; (*7*) Qadmous; (*8*) Wadi Jahannam; (*9*) Jdaida; (*10*) Mareen the East; (*11*) Bab Janneh; (*12*) Shuekha.

salinity of sea water acting upon the distribution of eury- or stenohaline forms that is especially well demonstrated by ostracodes; (6) the climatic regime in adjacent land areas and the distance of the sedimentation area from the provenance responsible for the composition and character of the terrigenous component; and (7) sedimentation rates that were substantially lower in epicontinental seas 0.001–0.005 mm/y than in geosynclinal basins (0.2–0.3 mm/y).

The studied Jurassic marine basins of the passive margin of the Mesozoic Tethys were located in the equatorial area. According to structural and tectonic features, the basins represented epicontinental shelf seas bounded in the south by the northern margin of the Arabian shield, and adjacent in the north to the northeastern Zagros. According to lithologic facies, the basins were in carbonate platform seas, shallow and rather remote from denudation areas, with outlines and dimensions relatively stable in the Middle Jurassic. Their depth and contours only slightly varied depending on transgressions and regressions of the Jurassic sea, which also determined the composition and character of deposits.

As for the bathymetric zonality, all zones from supralittoral to epibathyal were recorded in these basins. They were characterized by different foraminifer and ostracode assemblages.

The study of more than 90 sections of Jurassic sediments located in major mountains of Syria, namely, in the Anti-Lebanon, Coastal Mountains, Kurd-Dag, and Palmyrides (Figs. 1–3), permitted the establishment of a certain regularity in the distribution of foraminifer and ostracode assemblages in various facies. The taxonomic composition of these assemblages was dependent on the composition and character of rocks confined to different bathymetric zones of the basin (Fig. 4). Analyzing the relation between foraminifer and ostracode generic and specific composition and the bathymetric zones of the basin, we should make the following observations. Sediments of the deepest-water epibathyal and outer sublittoral zone yield rare foraminifers of Globigerinidae and Ceratobuliminidae families. The deposits reveal the presence of planktonic *Globuligerina* forms, which were widespread, though not numerous, in corresponding facies of the Bathonian and Callovian, and contain representatives of *Praelamarckina* and *Lamarckella* of the Ceratobuliminidae family. In this part of the basin ostracodes are represented by different species of *Cytherella* and *Bairdia* from Cytherellidae and Bairdiidae families. The corresponding sediments were recorded in the Aalenian and Bajocian of the Anti-Lebanon and in the Bajocian of the Coastal Mountains. They are represented by limestones, carbonate clays, and, to a lesser degree, by marls (Fig. 4).

The sediments that accumulated in the outer and middle sublittoral zone yield foraminifer and ostracode assemblages of different taxonomic composition. These deposits of the lower Middle Jurassic, namely, of the Bajocian and Bathonian, are located in the south of western Syria. Foraminifer assemblages from this basin zone are dominated by representatives of Lenticulinidae, Vaginulinidae, and Nodosariidae, namely, by *Lenticulina, Citharina,* and *Dentalina* genera (Fig. 5).

The depth reduction near the shoreline resulted in the increase of foraminifer taxonomic diversity, and sediments of the middle and partly inner sublittoral zone yield members of Polymorphinidae, Epistominidae, Prolixoplectidae, Pfenderinidae, and Crisalidinidae families. Bathonian and Callovian sediments of these bathymetric zones from the Anti-Lebanon and Coastal Mountains bear abundant species of *Redmondoides, Riuadhella, Eoguttulina,* and *Epistomina.*

Distribution patterns of ostracode assemblages in the outer, middle, and inner sublittoral zone somewhat differ from those of foraminifers. Ostracode taxonomic diversity in the outer sublittoral area is much lower than that of foraminifers. Their most diverse assemblages inhabited the middle and, especially, inner sublittoral zone.

In the outer sublittoral area, ostracode assemblages were impoverished and consisted only of *Cytherella* representatives from the Cytherellidae family. In the middle and inner sublittoral zone the ostracode fauna

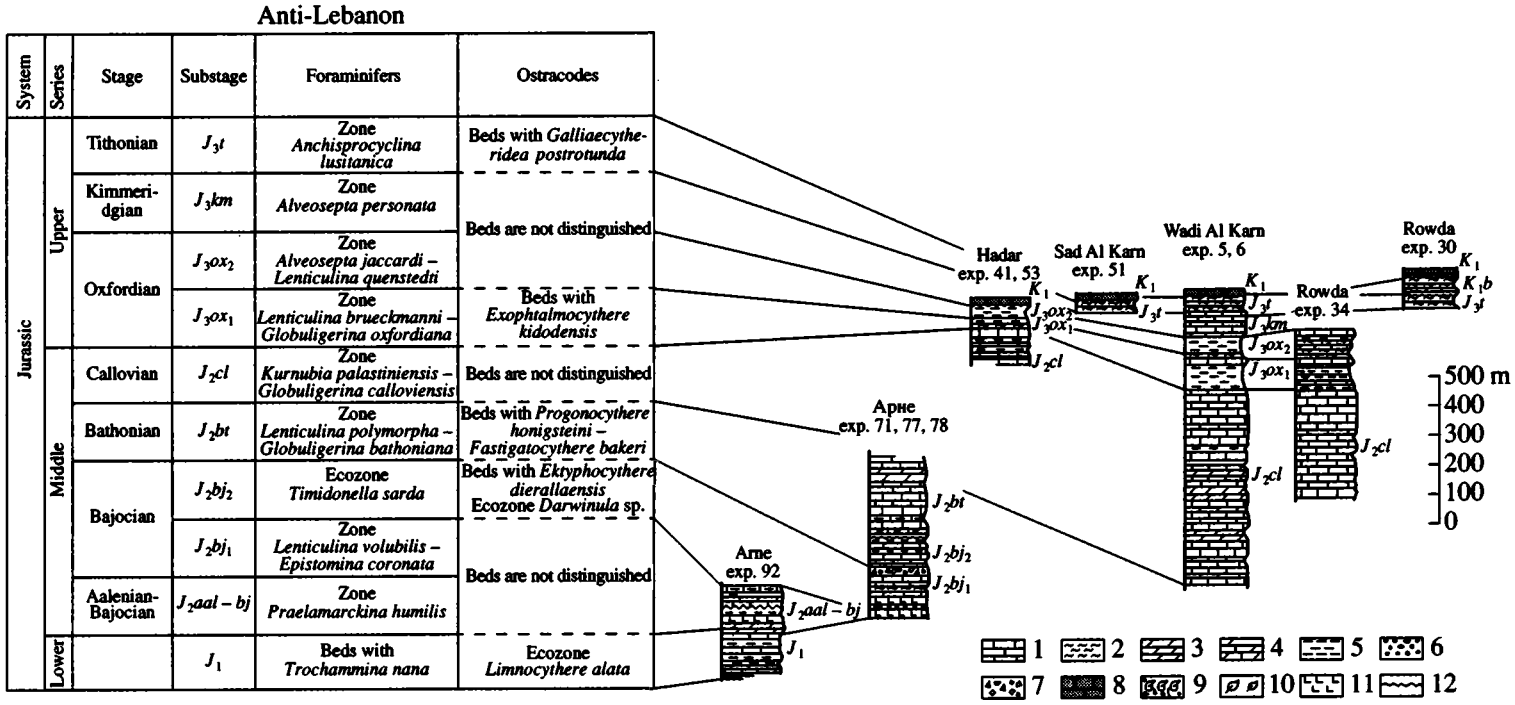

Fig. 2. Correlation of Jurassic sections in the Anti-Lebanon: 1, limestone; 2, marl; 3, dolomite; 4, dolomitic limestone; 5, clay; 6, conglomerate; 7, breccia; 8, sandstone; 9, macrofaunal remains; 10, plant remains; 11, volcanogenic rocks; 12, erosional boundary.

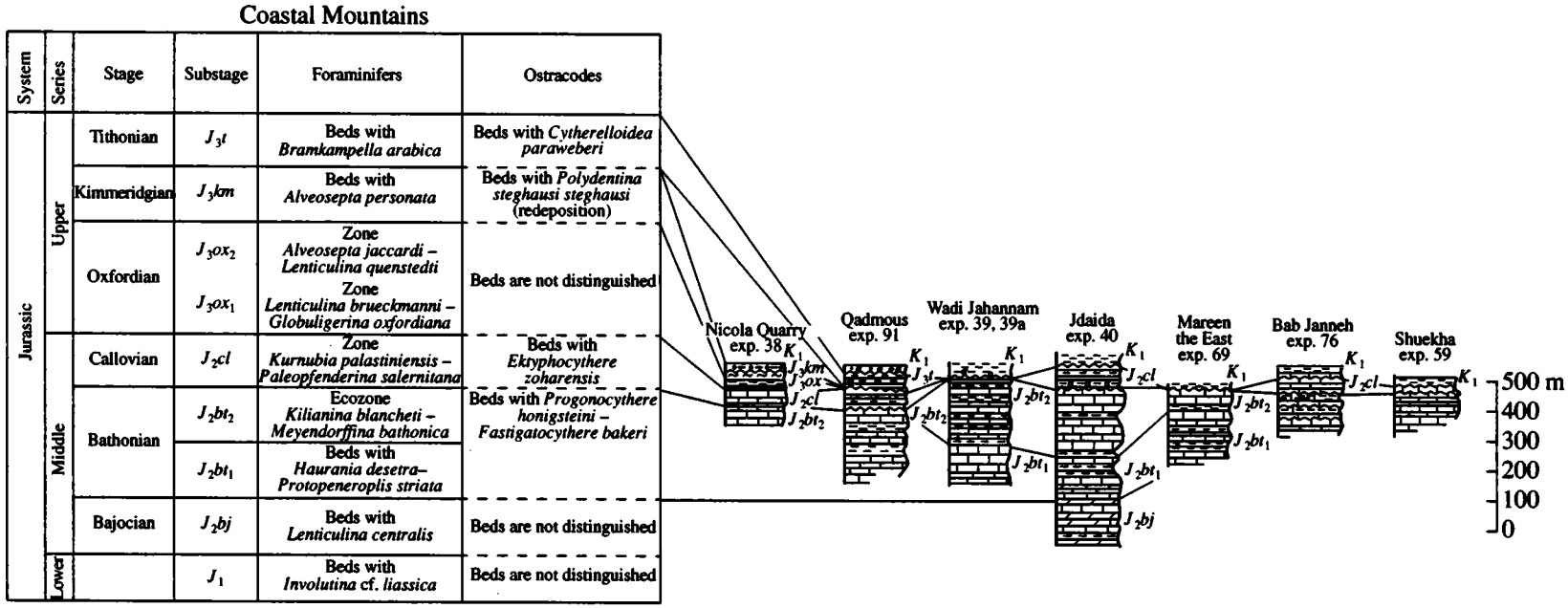

Fig. 3. Correlation of Jurassic sections from the Coastal Mountains (for symbols, see Fig. 2).

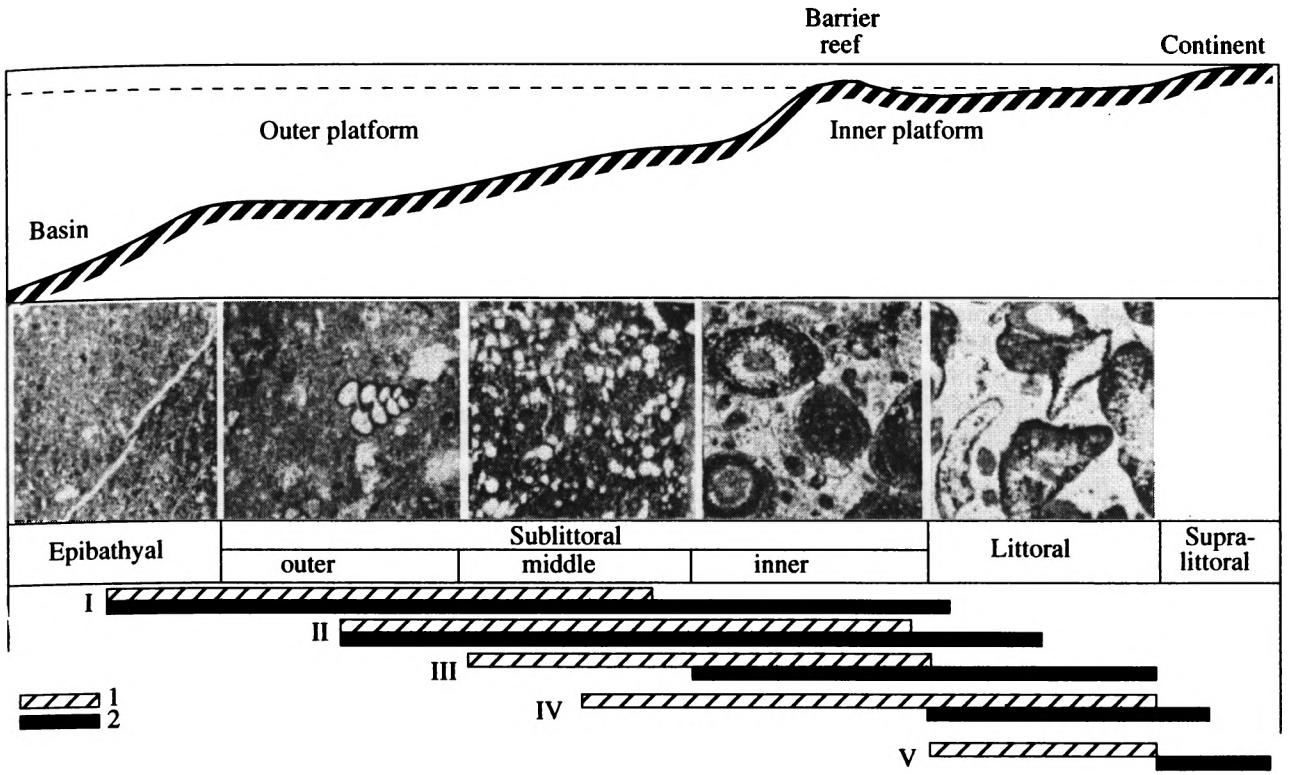

Fig. 4. Schematic model of the basin, distribution of foraminifer and ostracode species, and sedimentary facies in bathymetric zones in the Middle Jurassic basin of Syria: 1, foraminifers; 2, ostracodes; I–V, foraminifer and ostracode assemblages from different bathymetric zones (see Fig. 5 for composition).

was more diverse. It included mainly *Progonocythere, Procytheridea, Fastigatocythere, Glyptocythere, Glyptogatocythere,* and *Ektyphocythere* genera of the Progonocytheridae family, as well as species of *Cytherella* from the Cytherellidae family.

The outer sublittoral ostracode assemblage from the Coastal Mountains is somewhat more diverse compared to that of the Anti-Lebanon. It includes *Bairdia, Cytherella, Macrocypris,* and *Bythocypris* genera. Middle and inner sublittoral assemblages consist of species of *Glyptocythere, Glyptogatocythere, Fastigatocythere, Rectocythere, Macrodentina, Procytheridea, Progonocytheridea, Praeschuleridea, Bairdia, Cytherella,* and *Paracypris* genera.

The decrease of depth affected not only the taxonomic composition of benthic faunal assemblages, but the character of deposits as well. The latter were represented by organoclastic, oolitic, and pseudoolitic limestones that alternated with rare beds of clay and clayey marls. Sand-size particles appeared in the terrigenous component, though in separate layers the silt-size material still prevailed. The composition of benthic fauna was substantially enriched with widespread bivalve mollusks, brachiopods, gastropods, and with solitary and colonial corals.

The sediments of this sort are characteristic of the Bathonian and Callovian of the Coastal Mountains and, partly, of the Callovian in northernmost sections of the Anti-Lebanon.

The middle and inner sublittoral, as well as littoral shallow-water sediments of the Callovian and early Bathonian are distributed in the Coastal Mountains and Kurd-Dag. Organoclastic, sometimes coarse detrital limestones yield a peculiar foraminifer assemblage dominated by representatives of the Pfenderinidae and Cyclamminidae families, namely, of *Paleopfenderina, Pfenderella, Steinekella, Kurnubia,* and *Dhrumella* and *Flabellamminopsis* genera, respectively.

The most shallow-water Middle Jurassic deposits, partly characteristic of the inner sublittoral and littoral zones, are widespread enough. They were recorded in the Anti-Lebanon, Coastal Mountains, Kurd-Dag, and Palmyrides, where they accumulated in marginal parts of the basin in the late Bajocian and Bathonian. Foraminifer assemblages include here *Meyendorffina, Haurania, Kilianina, Limognella, Alzonella,* and *Timidonella* genera from Dictyoconidae, Hauraniidae, and Hottingeritidae families. Sediments with these foraminifer and ostracode assemblages were formed in the shallow zone of a carbonate platform sea. They are represented by coarse detrital, oolitic, and stromatoporoid limestones, as well as by algal bioherms.

Ostracode assemblages from the littoral zone differ in their structure from those of the middle and inner

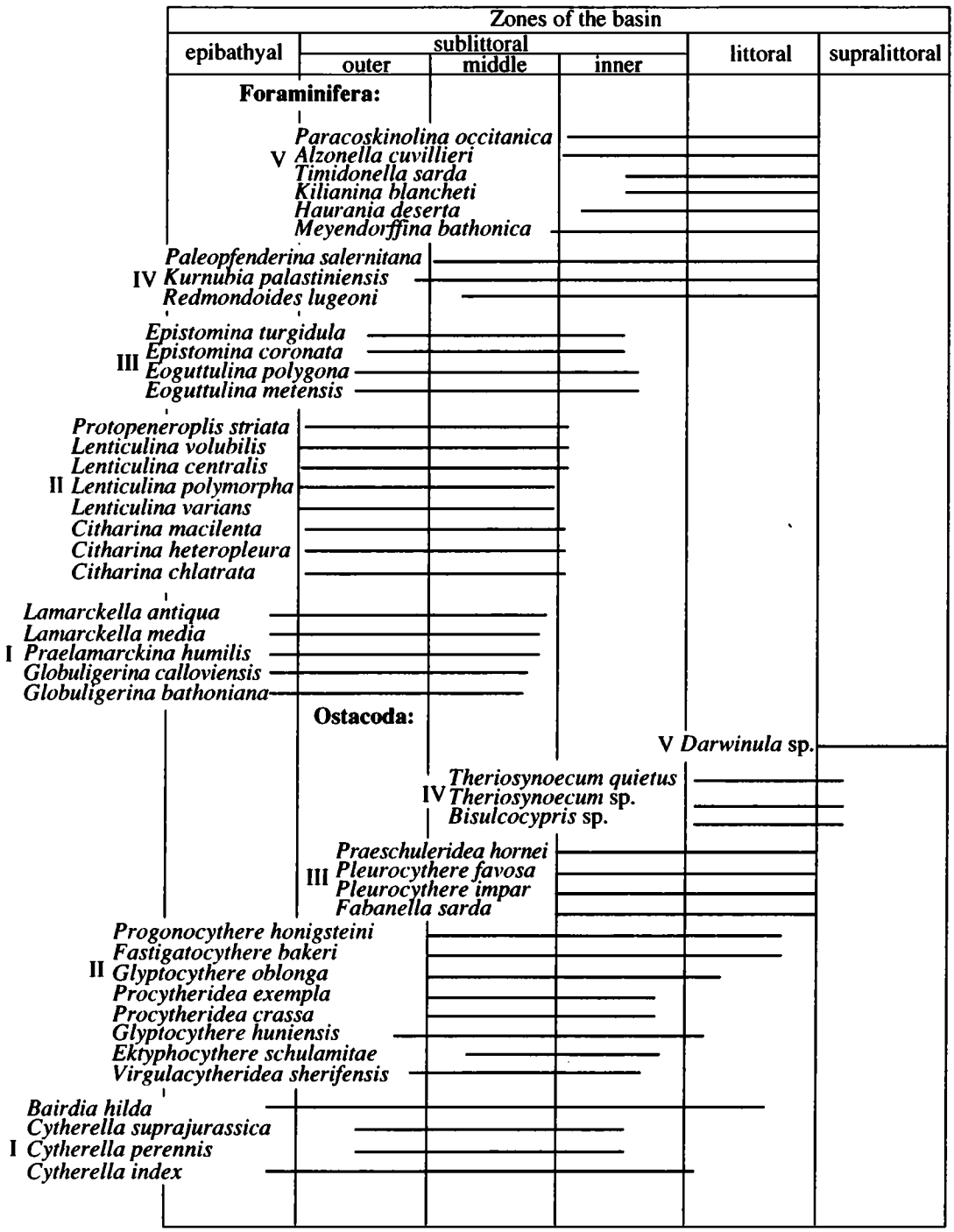

Fig. 5. Foraminifer and ostracode distribution in bathymetric zones of Middle Jurassic basins in Syria: I–V, foraminifer and ostracode assemblages in different bathymetric zones (see Fig. 4 for bathyal distribution).

sublittoral area. Very often they represent monogeneric assemblages with the predominance of a certain genus most adapted to the peculiar facial environment.

In the Anti-Lebanon, ostracode assemblages from the littoral zone include species of *Glyptocythere, Ektyphocythere,* and *Zergacythere* genera of the Progonocytheridea family, the *Schuleridea* genus of the Cytherideidae family, and *Cytherelloidea* genus of the Cytherelloidae family. In the Coastal Mountains, ostracodes of this bathymetric zone are represented by genera *Pleurocythere, Praeschuleridea, Fabanella, Theriosynoecum, Bisulcocypris, Paracypris, Schuleridea, Klieana,* and *Cytherelloidea.*

These data provide only a general idea of the distribution of foraminifer and ostracode assemblages in different parts of the basin and deserve further comment.

Our conclusions on the confinement of certain foraminifer assemblages to different bathymetric zones are based on the following: the analyzed lithologic composition of rocks and successive facial changes (Wilson, 1980; Sartorio and Venturini, 1988); the facial and bathymetric distribution of foraminifers in the Jurassic of southwestern France (Pelissie *et al.*, 1984) and Morocco (Hottinger, 1967; Sossipatrova and Rahhali, 1974); the analyzed paleoecology, sedimentology, and foraminifer distribution in the carbonate platform of the Tethys passive margin (Arnaud-Vanneau, 1980); the evolution of carbonate accumulation in continental margins (Konyukhov, 1987); and paleogeography of the Tethys (*Istoriya ...*, 1987; *Atlas ...*, 1993).

The study of Jurassic ostracodes from the Eastern Mediterranean has a long history and is related with a number of researchers from Europe, Israel, and Jordan (Oertli, 1966; Malz *et al.*, 1985; Depeche *et al.*,1987; Rosenfeld *et al.*, 1987; Rosenfeld and Honigstein, 1991; Rosenfeld, Oertli, *et al.*, 1987; Basha, 1980, 1992). Based on these and more general papers (Pokorny, 1978; Neale, 1988) on the paleoecologic distribution of different ostracode types, we analyzed assemblages for their confinement to certain bathymetric zones of the basin.

Let us consider the foraminifer and ostracode distribution in different bathymetric zones of the basin over the Bathonian. This time period corresponded to the expansion of transgression over the main part of Syria from the west to the Palmyrides. The wide distribution of Bathonian sediments makes it possible to trace most thoroughly the facies changes from deeper-water to near-shore shallow-water parts of the basin. The outer sublittoral and conceivably partly epibathyal sediments, represented by aphanic dense limestones without terrigenous components, compose the lower Bathonian sequence in the East Mareen, Wadi Jahannam, and Wadi Shkeir sections in the Coastal Mountains. The foraminifer assemblage is not taxonomically diverse, and the number of certain species is low. The assemblage includes *Globuligerina bathoniana* (Pazdro), *Epistomina nuda* (Terq.), *E. turgidula* Kapt., *Paulina paula* (Pazdro), *Globulina* sp., *Nautiloculina oolithica* Mohl., and rare *Riyadhella arabica* Redm.

The lower Bathonian sediments were also recorded in other Coastal Mountain sections, namely, in Jdaida, Shuekha, Qadmous, Bab Janneh, and Nicola Quarry. However, the early Bathonian foraminifers from these sections differ in composition from the assemblage mentioned above. The taxonomic diversity and species abundance are substantially higher. Many forms are represented by shells of different generations that indicate favorable environmental conditions. Shells are variable in dimensions and deprived of general morphologic peculiarities, such as inflation, flatness, ornamentation, and smoothness. The foraminifer assemblage consists of *Redmondoides reflexus* (Redm.), *R. medius* (Redm.), *R. lugeoni* (Sept.), *Paleopfenderina salernitana* (Sart. et Cresc.), *Kurnubia variabilis* Redm., and *Quinqueloculina* sp. The character of their life substratum considerably changes. In the studied sections, the lower Bathonian is represented by various fine detrital and clayey limestones. Separate layers and thin members of clay, clayey marls and limestones occur. The Bathonian foraminifer assemblage from the Anti-Lebanon has a different taxonomic composition. The facial changes in Bathonian sediments are distinct there in the section near the Arne Village. The beginning of the Bathonian was marked here by the general uplift, local hiatuses, and formation of near-shore littoral facies (coquina, coarse detrital limestones, and coarse sandstone beds with numerous erosion marks). Upward the section, the sediments give way to deeper-water oolitic, pseudoolitic, and fine detrital limestones, as well as to clayey limestones interbedded with clays and marls. We suggest that these deposits were formed in the middle and inner sublittoral area inhabited by the foraminifer assemblage that included *Lenticulina polymorpha* (Terq.), the index species of the corresponding zone, *L. centralis* (Terq.), *Astacolus filosus* (Terq.), *Eoguttulina triloba* (Terq.), *Lamarckella antiqua* Kapt., and *Globuligerina bathoniana* (Pazdro).

The late Bathonian time in the central West Syrian basin was characterized by a depth decrease toward the northeast and east and by the flourishing of foraminifer fauna. Many new genera and species appeared, while earlier forms continued to develop at that time, and the abundance and taxonomic diversity of assemblages considerably increased. Their most characteristic forms were *Dhrumella evoluta* Redm., *Meyendorffina bathonica* Auroze et Bizon, *Kilianina blancheti* Pfender, *Haurania deserta* Hens., *Amijiella amiji* (Hens.), *Limognella afaurei* Peissie et Peyb., *L. mesojurassica* Maync, and *Trocholina palastiniensis* Hens.

This is the most shallow-water foraminifer assemblage. Its distinct feature is the almost exclusive presence of Tethyan forms with a high degree of endemism at the level of genera and families.

Therefore, deep bathymetric zones of the Bathonian basin were inhabited by widespread cosmopolitan forms, namely, lenticulins, cytharins, epistomins, and partly planktonic foraminifers, whereas the role of endemic groups increased with the depth decrease and transition to sediments of a shallow carbonate platform. In the Middle Jurassic the endemic forms almost completely occupied the littoral and inner sublittoral area.

We discussed the distribution of Bathonian foraminifer assemblages in greater detail because sediments of this stratigraphic interval are better represented in sections of Syria. They are the most rich in fauna and most thoroughly studied.

Similar distribution patterns of foraminifers in different facies of successive bathymetric zones were

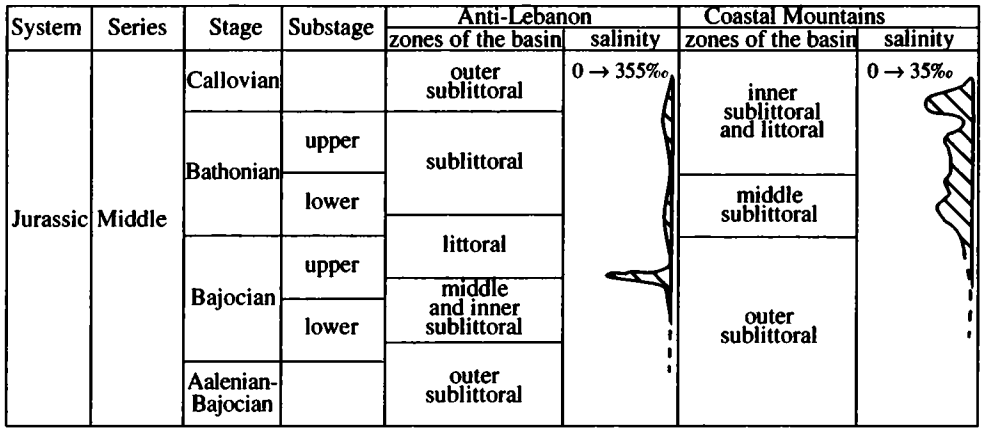

Fig. 6. Salinity fluctuations in bathymetric zones of Middle Jurassic basins in Syria.

recorded in the Bajocian and Callovian of Syria. The dependence of the assemblage composition and degree of endemism on facies bathymetric zones established for Bathonian foraminifer complexes was fully confirmed by the study of Callovian and Bajocian assemblages.

It is notable that outside Syria, in the peri-Tethyan basins, Crimea and Transcaucasia, as well as in the southern margin of the Russian platform (Dnieper–Don depression and northwestern Donetsk basin), Middle Jurassic sediments yielded foraminifers similar to assemblages I–III, which are typical of the outer and middle sublittoral zones of the Syrian basins. For example, in the eastern Crimea, namely, in Yanyshar Bay, the Meganom Peninsula, and Sudak synclinorium, the specific and generic composition of the Bajocian foraminifer assemblage is similar to that from the sections of the southern Anti-Lebanon (the Arne section) and Magara Ridge in the Sinai Peninsula. The reason is that precisely these zones of the basin with facies of pelitomorphic clayey or fine detrital limestones, were inhabited by cosmopolitan benthic and rare planktonic foraminifers. Diverse species of *Lenticulina, Citharina, Eoguttulina, Epistomina, Lamarckella,* and *Pseudolamarckina* genera were widespread in bottom sediments of high-latitude seas. According to these dominant groups, the peculiar boreal type of Jurassic foraminifer fauna was distinguished under the name of nodosariid–epistominid fauna in contrast to the cyclamminid–pfenderinid assemblage typical of the Tethys (Basov, 1974; Basov *et al.*, 1972; Basov, 1991).

The water salinity and its endurance or unsteadiness in different bathymetric zones and geological time spans are of significant interest for our understanding of foraminifer and especially ostracode distribution. As the investigation of ostracode assemblages showed, the relationship between steno- and euryhaline forms varied considerably in different parts of the Syrian basin. The determined general regularity is as follows: in the Middle Jurassic the deep, epibathyal and outer sublittoral zones were inhabited by ostracodes adapted to normal salinity, whereas shallow parts of the basin were dominated by ostracode genera and species known from brackish and freshwater basins (Fig. 6). In separate areas of the basin, for example, in the Anti-Lebanon (Arne section), the representatives of the freshwater *Darwinula* genus occurred in the Bajocian. Such, a paradoxical, at first glance, phenomenon as the presence of freshwater deposits in the marine sequence can, in our opinion, be a result of local freshening caused by freshwater in flows and seasonal heavy showers, which drastically changed the salinity in shallow parts of the basin.

Therefore, our investigations showed the following. Environmental conditions in Jurassic tropical shallow-water seas of the Tethyan passive margin were favorable for the development of benthic foraminifer and ostracode fauna.

The distribution of these benthic organisms in the Middle Jurassic seas was controlled by a combination of abiotic factors, of which the major one was the bathymetry. This factor was responsible for the character of the substratum inhabited by foraminifers and ostracodes, its composition and structure, and also for the distance from provenance, hydrodynamics of bottom water masses, and development of other benthic groups, which occurred together with foraminifers and ostracodes.

The highest taxonomic diversity of foraminifer and ostracode assemblages was recorded in sediments of the outer–middle and middle–inner sublittoral zones, respectively. In the littoral and inner sublittoral zones the diversity of foraminifers somewhat decreased. Ostracode assemblages became less diverse in extreme zones of the basin, namely, in its littoral and epibathyal areas. In the last bathymetric zone, the taxonomic diversity of foraminifers was substantially reduced as well.

We revealed a distinct differentiation of foraminifer and ostracode taxonomic composition across the bathymetric zones that they inhabited: in the epibathyal and outer–middle sublittoral zones, foraminifer assemblages were dominated by cosmopolitan forms, whereas near the shoals and shoreline, i.e., in the middle and inner sublittoral and littoral zones, the highly specialized Tethyan endemics prevailed.

The deep-water cosmopolitan nodosariid–epistominid foraminifer assemblages were more stable; their evolution was relatively slow and occurred mainly at the specific level. Tethyan cyclamminid–pfenderinid endemic faunas, which inhabited the near-shore shoals, developed more rapidly and at the higher generic level. Almost all 18 genera that appeared in the Bathonian became extinct by the end of the Callovian.

Ostracodes adapted to a wide range of salinity developed more rapidly than groups, that evolved under normal marine conditions. Freshwater ostracodes were represented by endemic groups of low taxonomic rank.

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The Jurassic–Cretaceous Boundary in the Boreal Realm (Biostratigraphy and Boreal–Tethyan Correlation)

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Abstract—A systematic analysis of the detailed biostratigraphy of Jurassic–Cretaceous boundary beds in principal regions of Russia is used to ground the correlation model of these deposits. The interregional correlations with respective strata in Europe and America demonstrate the range incompatibility of the terminal Jurassic stages accepted in the Boreal and Tethyan stratigraphic schemes, and suggest that the upper Volgian Substage should be considered part of the Cretaceous System. This means that the Jurassic–Cretaceous boundary in the Boreal Realm ought to be revised and to correspond, in our opinion, to the boundary between the lower and upper Volgian substages. The last level approaches, or is even identical to the accredited position of the Jurassic–Cretaceous boundary between the Tithonian and Berriasian stages.

Key words: *Jurassic, Cretaceous, Volgian Stage, Ryazanian Horizon, Berriasian, Tithonian, biostratigraphy, correlation*

The problem of the Jurassic–Cretaceous boundary is one of the principal problems in Mesozoic stratigraphy and was discussed many times during the last century. For a long time, this was due to poorly positioned boundary stages. According to the original definition of Oppel (1865), the Tithonian Stage included limestones, which actually yielded Berriasian fossils and were identified as the Berriasian Substage of the Valanginian (Coquand, 1871). Later, Renevier (1874) argued that this rock unit can be considered the basal stage of the Cretaceous System. However, to satisfy the priority principle, Toucas (1890) referred the Berriasian to the Jurassic System in contrast to Pictet (1867), the discoverer of fossils from the limestones in question. Many other researchers also considered this stage as a unit of the Cretaceous. Afterward, the range and validity of Tithonian and Berriasian stages, and, consequently, the position of the Jurassic–Cretaceous boundary have been repeatedly revised and discussed in many publications. The most recent among them are general reference works by M.S. Mesezhnikov (*Zony yurskoi ...*, 1982), N.I. Shul'gina (1985), and T.N. Bogdanova (*Zony melovoi ...*, 1989). It is also appropriate to mention here that, after a long discussion at the International Colloquium on the Jurassic–Cretaceous boundary at Lyon–Neuchatel in 1973, the majority of participants decided to consider the Berriasian as the lowermost stage of the Cretaceous, and to place the boundary between the Jurassic and Cretaceous systems at the base of this stage, or more precisely at the base of the *jacobi/grandis* Zone (*grandis* s.l. Zone in other terms). This scheme is now almost universally accepted.

However, the decision was ambiguous, because in the stratotype locality, rocks underlying the Berriasian deposits contain few ammonite remains (Cecca *et al.*, 1989), and the ammonite succession, which is necessary to establish the boundary of such a rank, cannot be distinguished here. Moreover, the type sections of the upper Tithonian Substage and Tithonian Stage as a whole never existed. The Tithonian Stage, as distinguished by Oppel, represented a synthetic unit; its stratotype was lacking, and only separate type sections from different localities were described as constituent parts of the stage. For instance, it was long believed that the upper Tithonian stratotype is represented by the Stramberg locality (Czechia). The mixed character of the Stramberg fauna was suspected long ago, however only detailed stratigraphic examination of this locality representing a series of limestone klippen demonstrated the absence of any clear ammonite sequence. The Stramberg ammonite fauna is distinctly mixed (Zeiss and Bachmayer, 1989), and its range corresponds to the interval from the middle Tithonian (the *fallauxi* Zone) up to and including the lower Berriasian (the *jacobi/grandis* Zone).

The most complete sections of the Tithonian and Berriasian stages, including the transitional beds, were discovered in the Betic Cordilleras, southeastern Spain, where they consist of marine deposits with abundant ammonites. The recent bed-by-bed examination of these sections and their ammonite fauna was successful in solving the problem of ammonite zonation in the Jurassic–Cretaceous boundary beds. In this region, Tavera (Oloriz and Tavera, 1982; Tavera, 1985) distinguished the *Simplisphinctes*, *Transitorius*, and *Durangites*

zones of the upper Tithonian, and the *jacobi, andrussowi*, and *boissieri* zones of the Berriasian. The latter correlate well with the Berriasian zones of the stratotype region in southeastern France. The large sample of data showed the expressive changes in ammonite fauna across the boundary between the *Durangites* and *jacobi* zones. This indicates the high stratigraphic rank of this level, and supports the decision of the Lyon–Neuchatel meeting to place the Jurassic–Cretaceous boundary of the Mediterranean scale between the *Durangites* and *jacobi/grandis* zones[1].

The substantiated boundary position in the Mediterranean scale (western Tethys) did not solve this particular problem as a whole. As is well-known, the Jurassic–Cretaceous transition is marked by a peak differentiation of marine fauna (including ammonites) that seems to be a response to several climatic and tectonic events. The sharp distinction between marine faunas of the indicated transitional time gave rise to parallel stages in stratigraphic practice: the Tithonian and Berriasian in the Tethyan regions, and the Volgian Stage and Ryazanian Horizon in the Boreal and Subboreal biochores.

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However in the 1960s and 1970s, some West European scientists (Casey, 1973; Kutek and Zeiss, 1974; Barthel, 1975) suggested that the Tithonian and Volgian stages are inadequate in their ranges and that the upper Volgian Substage may correspond to the Berriasian. There were adherents of this idea in Russia as well (Sasonova and Sasonov, 1979, 1983, 1984). Mesezhnikov (1989, and his contribution to *Zony yurskoi ...*, 1982) believed also that the uppermost unit of the Volgian Stage (the *Craspedites nodiger* Zone) is Cretaceous in age. As he noted, convincing arguments in favor of "sedative illusion" with respect to the congruence of the Tithonian and Volgian stages are lacking. Thus, despite the aforementioned consensus at the meeting in Lyon–Neuchatel, the problem of the Jurassic–Cretaceous boundary is still very controversial, and aspects such as the Boreal–Tethyan correlation and the level of Jurassic–Cretaceous boundary in the Boreal province are of primary importance.

[1] Nowadays, the lowermost Berriasian zone is often thought to correspond to the *Pseudosubplanites euxinus* Zone subdivided into the *jacobi* and *grandis* subzones (Zeiss, 1986; Cecca *et al.*, 1989).

Key regions for solving the aforementioned problems are those which comprise localities with mixed Boreal–Tethyan fauna. They are especially promising if fossils include the mixed ammonite taxa as in the Russian platform and northern Caucasus, or outside Russia, in Mangyshlak, the Crimea, and Poland. At the same time, it is reasonable to focus attention on regions with another type of fauna, where the Tethyan ammonites occur together with the Boreal bivalves of the *Buchia* group, because many researchers reported on the major significance of the latter for the Boreal–Tethyan correlation. In Russia, the interesting mixed faunas of the last type are known from the Jurassic–Cretaceous boundary beds of the Far East.

THE FAR EAST

In the Far East, the Upper Jurassic and Lower Cretaceous deposits bearing the Boreal buchiids are most widespread in the western Sea of Okhotsk region and southern Primor'e. For an extended period, we studied their sections bed-by-bed and meticulously examined ammonoids and buchiids (Sei and Kalacheva, 1990, 1992, 1993b).

The most complete sections of interest are in the Uda and Torom depressions (western Sea of Okhotsk region), where they include deposits of the Oxfordian, Kimmeridgian, Volgian, and Boreal Berriasian stages. Beginning with the lower Oxfordian, the entire bed sequence yielded samples with buchiids, which were used to establish the zonal scale with eight *Buchia* concurrent-range-zones, whose range mainly corresponds to the substage (Sei and Kalacheva, 1993b). This zonal sequence is similar to those established in North Siberia and the Russian platform, where *Buchia* scales were correlated with detailed ammonite zonations (Zakharov, 1981). The critical comparison with *Buchia* sequences in Siberia, northwestern Europe, northern Asia, and North America provided grounds for the reliable dating of our *Buchia* zones and was controlled by data on ammonites from the Sea of Okhotsk region.

Buchiids in this region are most abundant in the Volgian beds. We subdivided this stage into four zones: the lowermost *Buchia rugosa–B. mosquensis* Zone, whose range is compatible with the lower and partially middle substages of the Volgian stage; the *B. mosquensis–B. russiensis* and *B. russiensis–B. fischeriana* zones, which are in the middle Volgian Substage; and *B. piochii–B. terebratuloides* Zone corresponding to the upper Volgian Substage. Ammonites are very rare in this interval, however in the upper Volgian Substage (the *B. russiensis– B. fischeriana* Zone) there was an important find of *Durangites* sp. (Sei and Kalacheva, 1992, 1993b). As it was demonstrated above, the level demarcating the top of the *Durangites* Zone from the overlying *jacobi/grandis* Zone is identified as the Jurassic–Cretaceous boundary.

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The Lower Cretaceous deposits of the western Sea of Okhotsk region also contain *Buchia* remains including those of the *Buchia okensis* Zone referred to the basal Berriasian in the Boreal province. Unfortunately, the direct junction of *B. piochii–B. terebratuloides* and *B. okensis* zones has never been observed in the respective regional sections, and the boundary between the upper Volgian and Berriasian is rather arbitrary in this area.

In the southern Primor'e, the Upper Jurassic System is represented only by its terminal stage, whose deposits, in contrast to the Sea of Okhotsk region, contain abundant ammonites, but are depleted in *Buchia* remains. The ammonite assemblages here consist of distinct Tethyan forms, and using this fact we attributed their host deposits to the Tithonian Stage. In the lower part of its sequence, we distinguished the *Virgatosphinctes* cf. *mexicanus* Beds, whose fossils resemble the lower Tithonian fauna from Argentina (Leanza, 1980; Sei and Kalacheva, 1992). The ammonite assemblages with abundant haploceratids and oppelids, and diverse perisphinctids (*Subplanitoides, Aulacosphinctoides, Parapallasiceras, Lemencia, Sublithacoceras*, and *Aulacosphinctes*) correspond here to a higher level, which is subdivided in the lower *Pseudolissoceras zitteli* and upper *Aulacosphinctes proximus* zones spanning the main part of the middle Tithonian Substage. Both zonal assemblages contain some index species typical of middle Tithonian fauna from Argentina (Leanza, 1980), nonetheless their dominant forms belong to specific ammonite genera of the European type, and in general the assemblages are most similar to coeval faunas from southern Europe. Some of the European forms, for instance, *Semiformiceras* ex. gr. *semiforme* afford the opportunity to correlate the middle Tithonian zones of southern Primor'e with the *semiforme, fallauxi*, and perhaps *ponti* zones of Europe (Enay and Geyssant, 1975; Zeiss, 1968). Rare buchiids, which associate with the middle Tithonian ammonites of Primor'e, include *Buchia rugosa* (Fisch.) and *B. mosquensis* (Buch), whose assemblage corresponds to the respective zone of the standard *Buchia* zonation. In other words, the middle Tithonian deposits of Primor'e may be correlated with the lower and presumably basal middle Volgian substages of the Boreal scale.

The Upper Jurassic of southern Primor'e may also include the thick sandstone sequence occurring a distance apart from the beds described above. All along the sequence, the sandstones yielded buchiids (Sei and Kalacheva, 1990), among them the abundant *Buchia piochii* (Gabb), *B. terebratuloides* (Lah.) and *B. unschensis* (Pavl.) associated with rare *B. fischeriana* (Orb.) and *B. trigonoides* (Lah.), and single *B. volgensis* (Lah.) and *B.* aff. *okensis* (Pavl.). This assemblage is indicative of the upper Volgian *B. piochii–B. terebratuloides* Zone of the *Buchia* scale.

In addition to Boreal buchiids, these deposits yielded Tethyan ammonites useful for parallel dating. *Dalmasiceras* sp. from the middle part of the section is characteristic of the lower Berriasian in southern Europe, where it is most abundant in the basal *jacobi/grandis* Zone of the substage (Le Hegarat, 1973; Enay and Geyssant, 1975; Tavera, 1985). In upper beds, we found *Pseudosubplanites* cf. *grandis* (Maz.), *P.* aff. *combesi* Le Hegarat, *P.* sp., *Berriasella* ex gr. *jacobi* Maz., and *B.* sp. (Sei and Kalacheva, 1990) also corresponding to index species and the associated lower Berriasian forms from the European *jacobi/grandis* Zone. Using this similarity, we may reliably allocate ammonites from southern Primor'e in standard Jurassic and Cretaceous units of the European ammonite zonation. The resulting data show that the deposits in question correspond to either the lower Berriasian of the Tethyan scale, or the upper Volgian of the Boreal (*Buchia*) scale.

Mixed faunas of this type (associated Boreal buchiids and Tethyan ammonoids) are known in western North America (California, Oregon, and British Columbia). From the viewpoint of Boreal–Tethyan correlation, these faunas were discussed by Zeiss (1979), Jeletzky (1984), and Hoedemaeker (1987), whose inferences are comparable with conclusions we derived from studied sections in the Far East.

In California (Jones *et al.*, 1969), the upper part of the *Buchia russiensis* (*B. elderensis* after R. Imlay) Zone seems to be identical to the middle Volgian *B. russiensis–B. fischeriana* Zone in the Far East. It yielded the *Durangites–Kossmatia* ammonite assemblage and can be correlated with the upper Tithonian *Durangites* Zone as in the Far East regions. The zone underlies *B. fischeriana* and *B.* aff. *okensis* zones in California, and *B. piochii, B. fischeriana*, and *B. terebratuloides* zones in British Columbia. We consider all these zones as a summary equivalent of the upper Volgian *B. piochii–B. terebratuloides* Zone in the Far East. Tethyan *Spiticeras* and *Proniceras* from the lower–middle part of this summary interval may occur together only in the lower Berriasian *jacobi/grandis* Zone. The upper portion of the interval yielded *Paradontoceras* and *Substeuroceras*, and thus their host beds apparently correspond to the *Substeuroceras koeneni* Zone in Argentina (Leanza, 1980). At present, the last one is accepted as approximately corresponding to the *jacobi/grandis* and *occitanica* zones of the European scale (Zeiss, 1986; Tavera *et al.*, 1986).

Finally, the uppermost *B. okensis* and *B. uncitoides* zones of British Columbia along with analogous beds in California (*B. uncitoides* Zone) are comparable with the *Argentiniceras noduliferum* Zone of Argentina in the greatest part of their interval, which yielded *Neocosmoceras* and *Argentiniceras* forms. Accordingly in the European scale, they can be correlated with the *paramimounum* and possibly *picteti* subzones of the upper Berriasian *boissieri* Zone (Zeiss, 1986; Tavera *et al.*, 1986; Hoedemaeker, 1987). However, the

indicated ammonites occur in the middle and upper parts of the *Buchia* zones in question, therefore the entire *B. okensis–B. uncitoides* interval most likely corresponds to the whole *boissieri* Zone.

Thus, our analysis of mixed buchiid–ammonoid faunas from Jurassic–Cretaceous boundary beds in the Russian Far East and western North America allows us to state that Volgian and Tithonian stages are dissimilar in their stratigraphic ranges. The middle and upper Volgian substages correspond to the upper Tithonian and lower Berriasian substages respectively, when we consider the latter as an equivalent of the *jacobi/grandis* and *occitanica* zones. The *B. okensis* and *B. uncitoides* zones of the *Buchia* zonal scale, which are almost identical to the Boreal Berriasian, may be equivalent to the upper Berriasian *boissieri* Zone of the European scale.

NORTHERN CAUCASUS

The northern Caucasus is a key region for solving problems of the Boreal–Tethyan correlation and the position of the Jurassic–Cretaceous boundary. The Tithonian and Berriasian stages here comprise marine terrigenous–carbonate deposits, which are well exposed and contain ammonites most abundant in the Berriasian beds. After extensive study of these deposits and their ammonite fauna, Sacharov (1975, 1976, 1979, 1984; Sacharov *et al.*, 1987) elaborated the Berriasian ammonite zonation often used in interregional correlation practice. Unfortunately, this zonation is based on rather poor paleontological materials.

One of the most complete and well-exposed Berriasian sequences, containing abundant ammonites and bivalves, is that of the Urukh River (northern Osetia). In 1986–1987, we studied it as members of a scientific team headed by Prof. M.S. Mesezhnikov. The bed-by-bed sampling of micro- and macrofossils was an important part of this work. After the untimely demise of Mesezhnikov, we comprehensively investigated the collected ammonites and some bivalves (buchiids), and then suggested a new version of Berriasian zonation in the northern Caucasus.

Figure 1 shows the generalized scheme of a very detailed Urukh sequence of 109 beds that we identified together with Mesezhnikov and other colleagues. The 90-m-thick Berriasian deposits underlying Valanginian limestones rest here on the eroded surface of another limestone bed, which contains, according to the data of S.F. Makar'eva, the upper Tithonian calpionellids.

We believe that the very complete Urukh section, which is extremely rich in fossils, should be accepted as the reference Berriasian sequence for the northern Caucasus. Sacharov initially attributed this status to the section of the Myag-Sekyabir Canyon (Assa River basin in Ingushetia), where he studied in detail and described many guide fossils from the stratigraphic interval in question (Sacharov, 1975, 1976, 1984). However, an irregular fauna distribution in the somewhat inaccessible Myag-Sekyabir locality leads us to prefer the Urukh counterpart, although some stratigraphic levels in the former yielded very important paleontological materials. In general, both sections are similar in their structure, lithology, rock thickness, and diversity of faunal remains. We actually used data from both sections in our scheme of Berriasian biostratigraphy.

The ammonite assemblage with dominant large (up to 240 mm in diameter) *Mazenoticeras malbosiforme* Le Heg. is lowermost in the Urukh section. In French sequences, this species appears at the base of the *Tirnovella occitanica* Zone, i.e., in its *T. subalpina* Subzone (Le Hegarat, 1973). Other species of the assemblage do not contradict, or even support, this age. They encompass abundant *Jabronella* forms, including *J. incomposita* (Ret.) and *J. discrepans* (Ret.). Retowski (1893) described both species (the second one as a variety of the first form) from the "Feodosiya Marl" in the Crimea, but did not properly allocate them in the section. Bogdanova *et al.* (1984), who studied the Feodosiya locality in detail, concluded that *Jabronella* species are lacking in the very representative assemblage recovered from the lowermost Berriasian *Pseudosubplanites ponticus–P. grandis* Zone. Retowski apparently described the aforementioned forms from another level now inaccessible to observation. Le Hegarat reported on *Jabronella discrepans* occurrence in the lower *paramimounum* Subzone of the *Fauriella boissieri* Zone of the French sections, while Nikolov (1982) observed this form even in the *callisto* Zone of Bulgaria. According to Le Hegarat's data, genus *Jabronella* is widespread in the upper Berriasian *boissieri* Zone, and only a few species of the genus appear in the *occitanica* Zone. Thus, judging from the species list reported by Retowski, this ammonite assemblage should be characteristic of beds that are not younger than the *occitanica* Zone.

We detected *Retowskiceras* cf. *andrussowi* (Ret.) in basal beds of the Urukh section. This form is known as an element of zonal assemblage typical of the *P. ponticus–P. grandis* Zone in the Crimea (Bogdanova *et al.*, 1984). At the same time, it is accredited as the index species of the *andrussowi* Zone in southern Spain, which is considered identical to the *occitanica* Zone of French sections (Tavera, 1985). In other words, the stratigraphic range of this species is wide, and spans the entire interval of the lower Berriasian[2]. Among smaller ammonites we identified *Fauriella* cf. *clareti*, a species known from the *subalpina* and *privasensis* subzones of the *occitanica* Zone in southeastern France. In addition, this assemblage includes such specific forms as

[2] At present, the uniform subdivision of Berriasian into substages is absent. In the northern Caucasus, Sacharov (1979; Sacharov *et al.*, 1987) divided this stage into two substages on the basis of changes in ammonite fauna composition and lithology. His scheme shows the lower and upper substages as adequate to the summary interval of the *jacobi/grandis* and *occitanica* zones, and the *boissieri* Zone respectively (local analogues are included in both cases). In our opinion, this is well substantiated.

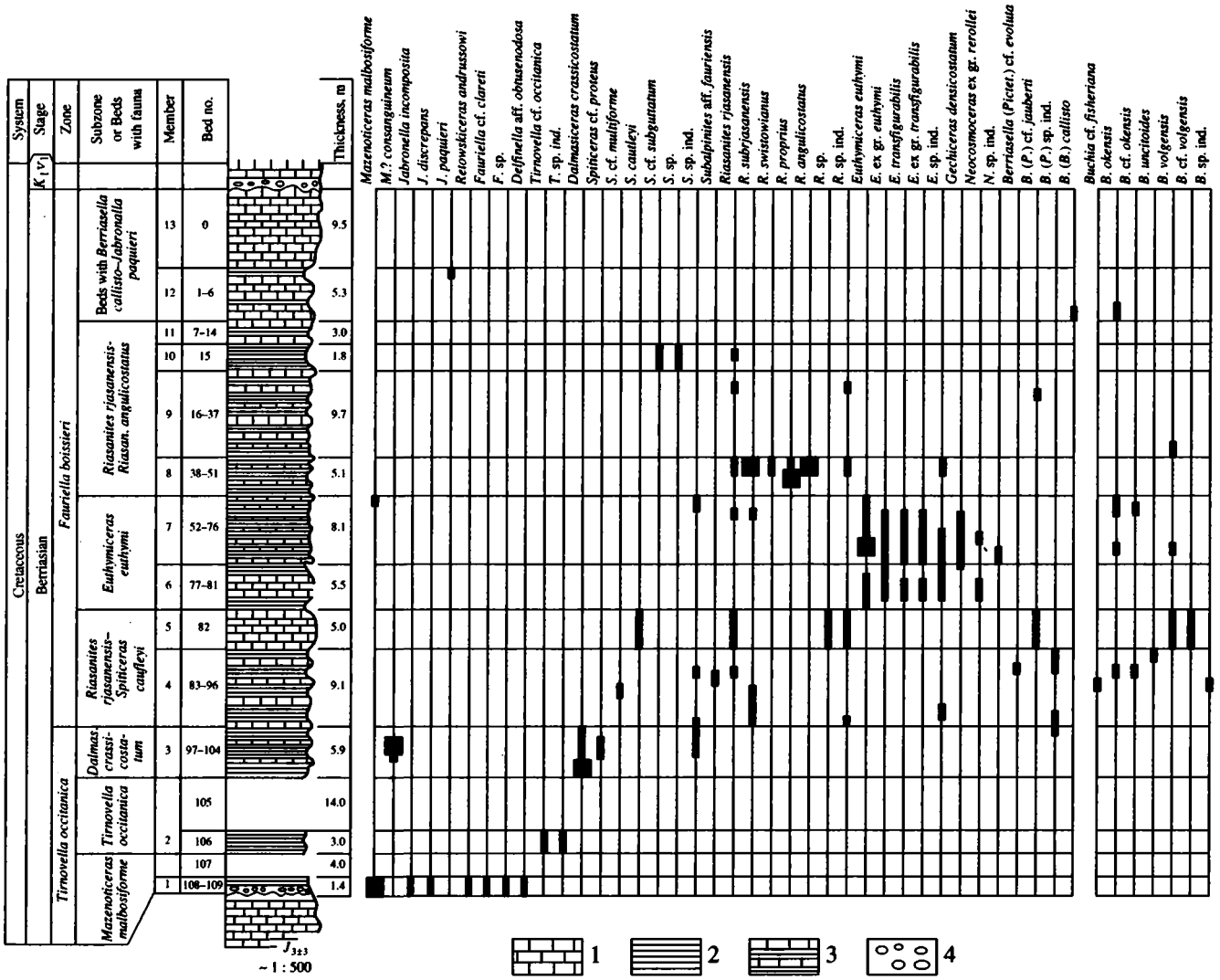

Fig. 1. Section of Berriasian deposits in the Urukh River basin (northern Caucasus) with indicated distribution ranges of ammonites and buchiids: 1, limestones; 2, clays, clayey marls, and siltstones; 3, fine alternation of limestone, clay, and marl; 4, breccia zones.

Delphinella aff. *obtusenodosa*. As is evident from the French sections, the coexistence of *Jabronella* and *Delphinella* forms seems to be most appropriate for the lower part of the *occitanica* Zone.

Thus, we may conclude that this assemblage corresponds to the lower part of the *occitanica* Zone, and most likely to the *subalpina* Subzone of the stratotype section. This conclusion does not fit the age determination suggested by Sacharov *et al.* (1987) for the basal beds in the Urukh locality. These researchers published an ammonite list, including diverse forms of *Pseudosubplanites, Fauriella*, and *Delphinella* genera associated with *Malbosiceras malbosi* (Pict.), and considered them as having originated from the basal Berriasian *Pseudosubplanites ponticus* Zone. In our collection, species of the *Pseudosubplanites* type are missing. Moreover, Sacharov (1976) reported that large *Malbosiceras malbosi* and *M.* spp. occur in basal beds of the Myag-Sekyabir locality. One *M. malbosi* specimen has been pictured (Sacharov, 1984; Plate V, Fig. 2); in our opinion, it is identical to *Mazenoticeras malbosiforme* from the Urukh locality. It is also important that *M. malbosi* is a characteristic element of the upper Berriasian ammonite assemblages, and we accordingly believe that basal Berriasian beds are missing in both the Urukh and Myag-Sekyabir localities. Kvantaliani (1989), who also studied the Urukh section, arrived at the same conclusion.

As Sacharov (1979, 1984) reported, the most representative assemblage of early Berriasian ammonites occurs in sections of the Gizel'don and Fiagdon rivers, where it includes *Pseudosubplanites ponticus* (Ret.), *P. combesi* Le Heg., *Malbosiceras malbosi* (Pict.), *M. paramimounum* (Maz.), *Fauriella latecostata* (Kil.)

and other forms. In our opinion, the recovered *P. ponticus* (Sacharov, 1984; Plate V, Fig. 1) should be referred to *Berriasella privasensis* (Pict.). Presumably in the same Fiagdon locality, Khimshiashvili (1976; Plate V, Figs. 1–3) identified similar ammonite forms also as *B. privasensis*. V.L. Egoyan's data on occurrence of *P. ponticus* and *P. euxinus* in the northwestern termination of the Greater Caucasus are also in need of revision. Thus, the question of the occurrence of basal Berriasian in the northern Caucasus remains open for discussion.

The next interval (about 21 m) in the Urukh section is almost unexposed. However, in a small outcrop we met single and poorly preserved specimens of *Tirnovella* cf. *occitanica* (Pict.) and *T.* sp. indet. They suggest that the interval should be attributed to the *occitanica* Zone, very likely to its middle part as suggested by data on ammonite assemblages from underlying and overlying deposits.

The overlying beds yielded the reference ammonite assemblage, which is very useful for interregional correlations. It consists mainly of *Dalmasiceras crassicostatum* Djan. specimens (over 100 in number) associated with specific *Mazenoticeras? consanguineum* (Ret.) and rare *Spiticeras* forms. The *Dalmasiceras* level is well expressed in the Myag-Sekyabir section, and has also been traced in western Abkhazia and some regions of the Crimea. In our collection, respective ammonite specimens are mostly represented by the single, but extremely variable species *D. crassicostatum*, although Sacharov (1984) and Kvantaliani (1989) also identified different species of this genus in the Urukh and Myag-Sekyabir localities. For western Abkhazia (Psou River), Khimshiashvili (1976) published pictures of *D. sublaevis* Maz. and *D. gevreyi* (Kil.), whose morphology is well inside the variability limits we noted for *D. crassicostatum*. Pictured specimens from the *Dalmasiceras* level of eastern, central, and southwestern Crimea include *D. crassicostatum, D. punctatum, D.* aff. *crassicostatum, D.* ex gr. *subspiticeratoides* Djan., and *D.* sp. (Drushchits and Kudryavtsev, 1960; Kvantaliani, 1989; Bogdanova *et al.*, 1981). Unfortunately, the dominant *D. crassicostatum* and its stratigraphic level were not considered in the summarizing work of Le Hegarat (1973). Nonetheless, H. Mazenot reported that this species occurs together with *D. dalmasi* and *D. punctatum*, the index species for the *dalmasi* Subzone of the *occitanica* Zone in the European scale. In the Crimea, this age determination is supported by the parallel occurrence of *D. punctatum*. Accordingly, the *D. crassicostatum* Beds of the Crimea and Caucasus can be reliably correlated with the stratotype *dalmasi* Subzone.

Abundant and peculiar *Mazenoticeras? consanguineum* forms, which occur together with *D. crassicostatum* in the Urukh locality, were approximately correlated with the ammonite species described by Retowski in the Feodosiya Marl of the Crimea. As was noted, their position in the last section has not been properly established. It appears, however, to be inside the lower Berriasian interval.

Thus, for the Urukh section, we consider it possible to distinguish here the standard *Tirnovella occitanica* Zone in its whole range corresponding to the distribution interval of the *Mazenoticeras* assemblage and enclosing the *subalpina* Subzone at the base and the *Dalmasiceras* level above it. We detected this zonal index species in the middle part of the interval, where it was also noted by Sacharov *et al.* (1987). Sacharov (1975) reported as well on abundant *T. occitanica* from the Myag-Sekyabir locality. Despite the fact that its pictured specimen (Sacharov, 1984; Plate V, Fig. 3) is actually *D. crassicostatum*, we do not exclude the possibility that true *T. occitanica* forms may occur at this level. In addition, we are of the opinion that the recognized zone can be subdivided into the following units: a *Mazenoticeras malbosiforme* Subzone analogous to the standard *subalpina* Subzone, a tentative *T. occitanica* Subzone approximately corresponding to the *privasensis* Subzone, and a *D. crassicostatum* Subzone identical to the *dalmasi* Subzone. These units, with the exception of the *crassicostatum* Subzone, are correlated with standard subdivisions only in a schematic way.

Above the *Dalmasiceras* Beds of the Urukh locality (within the interval of beds 82–96), we discovered an ammonite assemblage of primary significance for the Boreal–Tethyan correlation. Its dominant forms are representatives of the genus *Riasanites*, such as *R. rjasanensis* (Nik.), *R. subrjasanensis* (Nik.), and large (more than 200 mm in diameter) *R.* sp. Similarly abundant are *Spiticeras cautleyi* (Opp.), *S.* cf. *multiforme* Djan., and *S.* sp. indet. They associate with rare specimens of *Berriasella* (*Picteticeras*) *evoluta* Le Heg., *B.* (*P.*) sp., *Subalpinites* aff. *fauriensis* Maz., and single *Euthymiceras* sp. indet. The ammonite assemblage coexists with Buchia cf. *fischeriana* (Orb.) and *B.* sp. indet. (cf. *volgensis*) in the lower part of the interval, and with *B. okensis* (Pavl.) and *B. uncitoides* (Pavl.) in its upper portion. The *Riasanites* assemblage directly succeeding the *Dalmasiceras* forms was discovered for the first time in the Caucasus and Crimea. Sacharov *et al.* (1987) were of the opinion that in the northern Caucasus the *Dalmasiceras* level immediately preceds beds with *Euthymiceras, Gechiceras*, and *Neocosmoceras* faunas. However in the Myag-Sekyabir locality, both levels are separated by an interval over 40 m in thickness and nearly barren of ammonites. In the Crimea, there was even an idea to unite both fauna types as specifying the indivisible *Dalmasiceras–Euthymiceras* Beds, however Bogdanova convincingly demonstrated that these taxa never occur together and are always separated by stratigraphic intervals of different thickness.

Beyond the position directly above the *Dalmasiceras* Beds corresponding to the *dalmasi* Subzone, the age of fauna under consideration is indicated by the first occurrence level of *Euthymiceras* forms, which are

known to occur in southeastern France only in the *paramimounum* Subzone of the *boissieri* Zone, i.e., at the base of the middle Berriasian (Le Hegarat, 1973). In Bulgaria, *Euthymiceras* species were recorded in the *paramimounum* and *picteti* subzones (Nikolov, 1982), although we cannot agree with some fauna determinations here. The *paramimounum* Subzone in Algeria also includes representatives of the *Euthymiceras* genus (Benest *et al.*, 1977). The *paramimounum* and *picteti* subzones define the distribution range of angusticostate ammonite species distinguished as *Berriasella* (*Picteticeras*) *evoluta* Le Heg. in southeastern France. All these data clearly mean that the fauna in question ("lower" *Riasanites* level in our terms) belongs to the *paramimounum* Subzone, and most likely to its lower part.

The subsequent interval of the section (beds 52–81) corresponds in time to the flourishing of the *Euthymiceras* genus. It yielded abundant *E. euthymi* (Pict.), *E.* ex gr. *euthymi, E. transfigurabilis* (Bogosl.), and *E.* ex gr. *transfigurabilis*, which are variable, but tightly interrelated species. They associate with *Gechiceras densicostatum* Sach. and peculiar *Neocosmoceras* ex gr. *rerollei* (Paq.). Single specimens of *Riasanites rjasanensis, R. subrjasanensis*, and *Mazenoticeras malbosiforme* occur in the upper part of the interval, while rare *Buchia okensis* and *B. volgensis* (Lah.) are dispersed throughout it. As noted in southeastern France, the association of *E. euthymi* and *Neocosmoceras rerollei* is confined to the middle of the *paramimounum* Subzone. The same association is typical of Algeria (Benest *et al.*, 1977) and Bulgaria (Nikolov, 1982). It was also recorded in the Crimea (Bogdanova *et al.*, 1981) and Mangyshlak (Luppov *et al.*, 1988), where it was correlated with the *paramimounum* Subzone by analogy with French sections. The described Urukh assemblage should be similarly correlated with this subzone.

Upward in the Urukh section, the *Euthymiceras* assemblage is replaced by nearly monotaxonic *Riasanites* fauna. These "upper" *Riasanites* Beds comprise *R. rjasanensis* (Nik.), *R. subrjasanensis* (Nik.) *R. swistowianus* (Nik.), *R. proprius* (= *Gechiceras proprium* Sach.), very frequent *R. angulicostatus* (=*Tauricoceras angulicostatum* K. et L.), and a single specimen of *Euthymiceras* sp. indet. We collected these ammonites in a narrow interval 2.7 m in thickness (beds 38–46) above the *Euthymiceras–Neocosmoceras* Beds. The higher 16.2-m-thick interval (including bed 7) yielded only single specimens of *Spiticeras* and *Picteticeras* forms, yet according to Sacharov (personal communication), *Riasanites rjasanensis* and *R.* sp. were also recorded here. At the same time, many specimens of *Riasanites rjasanensis, R. subrjasanensis, R. swistowianus*, and *R.* sp. were recovered from the upper 40-m-thick interval of the Myag-Sekyabir section (above the *Euthymiceras–Neocosmoceras* Beds; Sacharov, 1975), and this fauna may be characteristic as well of the Urukh section within the entire interval in question (beds 7–51). We define the age level of "upper" *Riasanites* Beds in the following manner. In the Myag-Sekyabir locality (Sacharov, 1984; Plate XII, Fig. 4), they occur together with abundant *Fauriella boissieri* (Pict.), the index species of the like-named zone, which are most abundant in the *picteti* Subzone (Le Hegarat, 1973). Above the main *Riasanites* level of the Urukh section, we found a *Berriasella* (*Picteticeras*) form similar to *B.* (*P.*) *jauberti* (Maz.) from the *picteti* Subzone. Finally, in both localities, basal beds of the interval yielded single *Euthymiceras* specimens. Taking into account that lower *Riasanites* and *Euthymiceras–Neocosmoceras* beds correspond to the *paramimounum* Subzone, we may conclude that upper *Riasanites* Beds partly correspond to its upper level, but mainly span the *picteti* Subzone. Bogdanova, who also correlated Caucasian sequences with the standard scale, arrived at nearly the same conclusion (*Zony melovoi* ..., 1989; *Zonal'naya* ..., 1991).

In the uppermost portion of the Urukh section (beds 0–6), we discovered *Jabronella paquieri* (Simon.) (bed 1), and Sacharov (1984; Plate VII, Fig. 5) reported on *Berriasella* (*B.*) cf. *callisto* (Orb.) presumably originating from bed 6. These species are known from both the *picteti* and *callisto* subzones, yet in our case they most likely characterize the latter, because one of the discovered ammonites is its index species. In the Myag-Sekyabir locality, the interval above the upper *Riasanites* Beds is insignificant in thickness and presumably incomplete.

The data presented here, especially the index species occurrence, are sufficient, in our opinion, to state that in the northern Caucasus, the upper Berriasian includes the standard *boissieri* Zone in its total range. This range above the *Dalmasiceras crassicostatum* Subzone (the equivalent of the *dalmasi* Subzone) is indicated by the first occurrence level of *Euthymiceras* species at the base, and by *Berriasella callisto* and *Jabronella paquieri* at the top. Bogdanova was first to assume in 1989 the presence of this zone in the Caucasus, and later in the entire southern region of the CIS (*Zonal'naya* ..., 1991). Based on recovered ammonite assemblages, we distinguish within the *boissieri* Zone the following subzones and beds (from the base upward): the *Riasanites rjasanensis–Spiticeras cautleyi* Subzone, the *Euthymiceras euthymi* Subzone, the *Riasanites rjasanensis–R. angulicostatus* Subzone, and the *Berriasella callisto–Jabronella paquieri* Beds. Two lower units are correlative with the stratotype *paramimounum* Subzone; the next one is almost identical to the *picteti* Subzone, and overlying beds correspond to the *callisto* Subzone. We indicated paired index species for two subzones because of the wide stratigraphic range of *R. rjasanensis*, which is very significant for correlations between the Boreal and Tethyan sequences. We also considered the possibility of outlining the *Riasanites rjasanensis* Zone of the Urukh section, whose interval may be comparable with that of the *boissieri* Zone. However, the new zone would be distinct from the *rjasanensis* Zone suggested by Sacharov

for the entire northern Caucasus, and they would both be deviant from the classical *R. rjasanensis* Zone of the Russian platform.

The detected zonation is well interrelated with the previous scheme suggested for the northern Caucasus (Sacharov *et al.*, 1987), because they both are based on the Urukh and Myag-Sekyabir sections. Nevertheless, our scheme excludes the basal Berriasian, and includes the lower *Riasanites* level. It correlates well with detailed Berriasian biostratigraphy of the Crimea and Mangyshlak. In contrast to the northern Caucasus, sections in the Crimea comprise the *Pseudosubplanites ponticus–P. grandis* Zone, which is paleontologically well substantiated as corresponding to the basal Berriasian *jacobi/grandis* Zone of the European scale (Bogdanova *et al.*, 1984). The reference level in both regions is represented by the *Dalmasiceras crassicostatum* Beds. Our *Euthymiceras–Neocosmoceras* and upper *Riasanites* beds correspond well to the *Tauriceras* Beds of the Crimea, whose index taxon is, in our opinion, the junior synonym of the *Riasanites* genus. Nonetheless, equivalents of our lower *Riasanites* Beds are still unknown in Crimean sections. The Berriasian sequence in Mangyshlak is considerably reduced and apparently corresponds only to a part of the upper substage (Luppov *et al.*, 1988). Local zonal units distinguished here are correlative with the *Euthymiceras euthymi* and *Riasanites rjasanensis–R. angulicostatus* subzones of the northern Caucasus.

Thus, with the exception of the basal level, the Berriasian sequence of the northern Caucasus can be considered as most representative for the southern regions of Russia and the CIS. Its ammonite assemblages include index species of standard zones and many taxa, which are important, on the one hand, for correlations with the standard European scale and, on the other hand, with Subboreal Lower Cretaceous sequences in the Russian platform.

RUSSIAN PLATFORM

The central Russian platform is a stratotype region of the Volgian Stage and Ryazanian Horizon representing the Jurassic–Cretaceous boundary units of the Boreal Realm. The Volgian Formation (Stage) was initially attributed to the Jurassic. This status remained unchanged, although subsequent stratigraphic revisions modified its range and subdivision scheme many times. The discovery of *Gravesia* forms at the base of Volgian beds in Subboreal and Boreal provinces of Russia made it possible to correlate the basal Tithonian and Volgian levels (*Zony yurskoi* ..., 1982; Mesezhnikov, 1989). As for the top levels of both stages, they were adopted as synchronous *a priori*. In all current stratigraphic schemes for the Russian territory, the Volgian Stage is terminal in the Jurassic System and considered as the Boreal equivalent of the Tithonian. Nevertheless, the idea to refer the upper Volgian Substage, or its part, to the Cretaceous System has long been known. We will argue for this viewpoint and consider the detailed correlation between the Tithonian and Volgian stages somewhat later, but first we will discuss the stratigraphic position of the Ryazanian Horizon, the basal unit of the Boreal Cretaceous.

At the moment of recognition, Bogoslowsky (1895) allocated the Ryazanian Horizon at the base of the Neocomian, and correlated it with the *"Hoplites boissieri"* Zone of Western Europe, which was accepted at that time as approximately corresponding in range to the Berriasian Stage. Inside the horizon, Bogoslowsky distinguished three units: lower beds with abundant *"Hoplites"* (= *Riasanites*) ex gr. *rjasanensis*; middle beds, where first *"Hoplites"* (= *Euthymiceras*) *transfigurabilis*, *"H."* (*E.*) *hospes*, and similar forms appearing in association with more rare *Riasanites* forms; and upper beds, where representatives of both groups are almost absent. Afterward, A.P. Pavlov, N.T. Sasonov, I.G. Sasonova, and P.A. Gerasimov subdivided the horizon into *Riasanites rjasanensis* and *Surites spasskensis* (or *S. tzikwinianus*) zones. Without any detailed analysis, the horizon was correlated with the *boissieri* Zone in its present interpretation, while the lower Berriasian was accepted as corresponding to the hiatus in sedimentation (Sachs *et al.*, 1975; Sasonova, 1977). This viewpoint is still universally accepted and expressed in principal resolutions (*Zony melovoi* ..., 1989; *Zonal'naya* ..., 1991).

In the 1970s, Mesezhnikov and his colleagues studied in detail the Ryazanian Horizon in the Oka River basin, where they recognized basal Ryazanian beds bearing such taxa as *Garniericeras, Craspedites*, and *Hectoroceras kochi* in addition to *Riasanites* and *Euthymiceras* forms. As a result, the *Riasanites rjasanensis* Zone was divided into the *Garniericeras, Hectoroceras kochi*, and *Euthymiceras transfigurabilis* beds capped by the *Surites tzikwinianus* Zone (Casey *et al.*, 1977; Mesezhnikov *et al.*, 1979). Elaborating this scheme in 1984, Mesezhnikov suggested a superdetailed zonation of the Ryazanian Horizon, which includes the following units (from the base upward): the *R. rjasanensis–Garniericeras subclypeiforme* Zone, the *R. rjasanensis–Hectoroceras kochi* Zone, the *R. rjasanensis–Surites spasskensis* Zone with three ammonite levels, and the *Surites tzikwinianus* Zone. Two newly introduced lower subdivisions were of crucial importance. *Garniericeras* and *Craspedites* forms from the horizon base are characteristic elements of the upper Volgian fauna, and confirm, in Mesezhnikov's opinion, the idea of N.A. Bogoslowsky and A.P. Pavlov that any significant hiatus between the Volgian and Ryazanian beds is missing. The erosion at the base of the Ryazanian Horizon is beyond any doubt; however, similar fauna from boundary beds suggests an insignificant time span for the respective hiatus in sedimentation.

Using these arguments, Mesezhnikov lowered the Ryazanian Horizon base down to the base of the *occi-*

known to occur in southeastern France only in the *paramimounum* Subzone of the *boissieri* Zone, i.e., at the base of the middle Berriasian (Le Hegarat, 1973). In Bulgaria, *Euthymiceras* species were recorded in the *paramimounum* and *picteti* subzones (Nikolov, 1982), although we cannot agree with some fauna determinations here. The *paramimounum* Subzone in Algeria also includes representatives of the *Euthymiceras* genus (Benest *et al.*, 1977). The *paramimounum* and *picteti* subzones define the distribution range of angusticostate ammonite species distinguished as *Berriasella* (*Picteticeras*) *evoluta* Le Heg. in southeastern France. All these data clearly mean that the fauna in question ("lower" *Riasanites* level in our terms) belongs to the *paramimounum* Subzone, and most likely to its lower part.

The subsequent interval of the section (beds 52–81) corresponds in time to the flourishing of the *Euthymiceras* genus. It yielded abundant *E. euthymi* (Pict.), *E.* ex gr. *euthymi, E. transfigurabilis* (Bogosl.), and *E.* ex gr. *transfigurabilis*, which are variable, but tightly interrelated species. They associate with *Gechiceras densicostatum* Sach. and peculiar *Neocosmoceras* ex gr. *rerollei* (Paq.). Single specimens of *Riasanites rjasanensis, R. subrjasanensis*, and *Mazenoticeras malbosiforme* occur in the upper part of the interval, while rare *Buchia okensis* and *B. volgensis* (Lah.) are dispersed throughout it. As noted in southeastern France, the association of *E. euthymi* and *Neocosmoceras rerollei* is confined to the middle of the *paramimounum* Subzone. The same association is typical of Algeria (Benest *et al.*, 1977) and Bulgaria (Nikolov, 1982). It was also recorded in the Crimea (Bogdanova *et al.*, 1981) and Mangyshlak (Luppov *et al.*, 1988), where it was correlated with the *paramimounum* Subzone by analogy with French sections. The described Urukh assemblage should be similarly correlated with this subzone.

Upward in the Urukh section, the *Euthymiceras* assemblage is replaced by nearly monotaxonic *Riasanites* fauna. These "upper" *Riasanites* Beds comprise *R. rjasanensis* (Nik.), *R. subrjasanensis* (Nik.) *R. swistowianus* (Nik.), *R. proprius* (= *Gechiceras proprium* Sach.), very frequent *R. angulicostatus* (=*Tauricoceras angulicostatum* K. et L.), and a single specimen of *Euthymiceras* sp. indet. We collected these ammonites in a narrow interval 2.7 m in thickness (beds 38–46) above the *Euthymiceras–Neocosmoceras* Beds. The higher 16.2-m-thick interval (including bed 7) yielded only single specimens of *Spiticeras* and *Picteticeras* forms, yet according to Sacharov (personal communication), *Riasanites rjasanensis* and *R.* sp. were also recorded here. At the same time, many specimens of *Riasanites rjasanensis, R. subrjasanensis, R. swistowianus*, and *R.* sp. were recovered from the upper 40-m-thick interval of the Myag-Sekyabir section (above the *Euthymiceras–Neocosmoceras* Beds; Sacharov, 1975), and this fauna may be characteristic as well of the Urukh section within the entire interval in question (beds 7–51). We define the age level of "upper" *Riasanites* Beds in the following manner. In the Myag-Sekyabir locality (Sacharov, 1984; Plate XII, Fig. 4), they occur together with abundant *Fauriella boissieri* (Pict.), the index species of the like-named zone, which are most abundant in the *picteti* Subzone (Le Hegarat, 1973). Above the main *Riasanites* level of the Urukh section, we found a *Berriasella* (*Picteticeras*) form similar to *B.* (*P.*) *jauberti* (Maz.) from the *picteti* Subzone. Finally, in both localities, basal beds of the interval yielded single *Euthymiceras* specimens. Taking into account that lower *Riasanites* and *Euthymiceras–Neocosmoceras* beds correspond to the *paramimounum* Subzone, we may conclude that upper *Riasanites* Beds partly correspond to its upper level, but mainly span the *picteti* Subzone. Bogdanova, who also correlated Caucasian sequences with the standard scale, arrived at nearly the same conclusion (*Zony melovoi* ..., 1989; *Zonal'naya* ..., 1991).

In the uppermost portion of the Urukh section (beds 0–6), we discovered *Jabronella paquieri* (Simon.) (bed 1), and Sacharov (1984; Plate VII, Fig. 5) reported on *Berriasella* (*B.*) cf. *callisto* (Orb.) presumably originating from bed 6. These species are known from both the *picteti* and *callisto* subzones, yet in our case they most likely characterize the latter, because one of the discovered ammonites is its index species. In the Myag-Sekyabir locality, the interval above the upper *Riasanites* Beds is insignificant in thickness and presumably incomplete.

The data presented here, especially the index species occurrence, are sufficient, in our opinion, to state that in the northern Caucasus, the upper Berriasian includes the standard *boissieri* Zone in its total range. This range above the *Dalmasiceras crassicostatum* Subzone (the equivalent of the *dalmasi* Subzone) is indicated by the first occurrence level of *Euthymiceras* species at the base, and by *Berriasella callisto* and *Jabronella paquieri* at the top. Bogdanova was first to assume in 1989 the presence of this zone in the Caucasus, and later in the entire southern region of the CIS (*Zonal'naya* ..., 1991). Based on recovered ammonite assemblages, we distinguish within the *boissieri* Zone the following subzones and beds (from the base upward): the *Riasanites rjasanensis–Spiticeras cautleyi* Subzone, the *Euthymiceras euthymi* Subzone, the *Riasanites rjasanensis–R. angulicostatus* Subzone, and the *Berriasella callisto–Jabronella paquieri* Beds. Two lower units are correlative with the stratotype *paramimounum* Subzone; the next one is almost identical to the *picteti* Subzone, and overlying beds correspond to the *callisto* Subzone. We indicated paired index species for two subzones because of the wide stratigraphic range of *R. rjasanensis*, which is very significant for correlations between the Boreal and Tethyan sequences. We also considered the possibility of outlining the *Riasanites rjasanensis* Zone of the Urukh section, whose interval may be comparable with that of the *boissieri* Zone. However, the new zone would be distinct from the *rjasanensis* Zone suggested by Sacharov

for the entire northern Caucasus, and they would both be deviant from the classical *R. rjasanensis* Zone of the Russian platform.

The detected zonation is well interrelated with the previous scheme suggested for the northern Caucasus (Sacharov *et al.*, 1987), because they both are based on the Urukh and Myag-Sekyabir sections. Nevertheless, our scheme excludes the basal Berriasian, and includes the lower *Riasanites* level. It correlates well with detailed Berriasian biostratigraphy of the Crimea and Mangyshlak. In contrast to the northern Caucasus, sections in the Crimea comprise the *Pseudosubplanites ponticus–P. grandis* Zone, which is paleontologically well substantiated as corresponding to the basal Berriasian *jacobi/grandis* Zone of the European scale (Bogdanova *et al.*, 1984). The reference level in both regions is represented by the *Dalmasiceras crassicostatum* Beds. Our *Euthymiceras–Neocosmoceras* and upper *Riasanites* beds correspond well to the *Tauriceras* Beds of the Crimea, whose index taxon is, in our opinion, the junior synonym of the *Riasanites* genus. Nonetheless, equivalents of our lower *Riasanites* Beds are still unknown in Crimean sections. The Berriasian sequence in Mangyshlak is considerably reduced and apparently corresponds only to a part of the upper substage (Luppov *et al.*, 1988). Local zonal units distinguished here are correlative with the *Euthymiceras euthymi* and *Riasanites rjasanensis–R. angulicostatus* subzones of the northern Caucasus.

Thus, with the exception of the basal level, the Berriasian sequence of the northern Caucasus can be considered as most representative for the southern regions of Russia and the CIS. Its ammonite assemblages include index species of standard zones and many taxa, which are important, on the one hand, for correlations with the standard European scale and, on the other hand, with Subboreal Lower Cretaceous sequences in the Russian platform.

RUSSIAN PLATFORM

The central Russian platform is a stratotype region of the Volgian Stage and Ryazanian Horizon representing the Jurassic–Cretaceous boundary units of the Boreal Realm. The Volgian Formation (Stage) was initially attributed to the Jurassic. This status remained unchanged, although subsequent stratigraphic revisions modified its range and subdivision scheme many times. The discovery of *Gravesia* forms at the base of Volgian beds in Subboreal and Boreal provinces of Russia made it possible to correlate the basal Tithonian and Volgian levels (*Zony yurskoi* ..., 1982; Mesezhnikov, 1989). As for the top levels of both stages, they were adopted as synchronous *a priori*. In all current stratigraphic schemes for the Russian territory, the Volgian Stage is terminal in the Jurassic System and considered as the Boreal equivalent of the Tithonian. Nevertheless, the idea to refer the upper Volgian Substage, or its part, to the Cretaceous System has long been known. We will argue for this viewpoint and consider the detailed correlation between the Tithonian and Volgian stages somewhat later, but first we will discuss the stratigraphic position of the Ryazanian Horizon, the basal unit of the Boreal Cretaceous.

At the moment of recognition, Bogoslowsky (1895) allocated the Ryazanian Horizon at the base of the Neocomian, and correlated it with the *"Hoplites boissieri"* Zone of Western Europe, which was accepted at that time as approximately corresponding in range to the Berriasian Stage. Inside the horizon, Bogoslowsky distinguished three units: lower beds with abundant *"Hoplites"* (= *Riasanites*) ex gr. *rjasanensis*; middle beds, where first *"Hoplites"* (= *Euthymiceras*) *transfigurabilis*, *"H."* (*E.*) *hospes*, and similar forms appearing in association with more rare *Riasanites* forms; and upper beds, where representatives of both groups are almost absent. Afterward, A.P. Pavlov, N.T. Sasonov, I.G. Sasonova, and P.A. Gerasimov subdivided the horizon into *Riasanites rjasanensis* and *Surites spasskensis* (or *S. tzikwinianus*) zones. Without any detailed analysis, the horizon was correlated with the *boissieri* Zone in its present interpretation, while the lower Berriasian was accepted as corresponding to the hiatus in sedimentation (Sachs *et al.*, 1975; Sasonova, 1977). This viewpoint is still universally accepted and expressed in principal resolutions (*Zony melovoi* ..., 1989; *Zonal'naya* ..., 1991).

In the 1970s, Mesezhnikov and his colleagues studied in detail the Ryazanian Horizon in the Oka River basin, where they recognized basal Ryazanian beds bearing such taxa as *Garniericeras, Craspedites*, and *Hectoroceras kochi* in addition to *Riasanites* and *Euthymiceras* forms. As a result, the *Riasanites rjasanensis* Zone was divided into the *Garniericeras, Hectoroceras kochi*, and *Euthymiceras transfigurabilis* beds capped by the *Surites tzikwinianus* Zone (Casey *et al.*, 1977; Mesezhnikov *et al.*, 1979). Elaborating this scheme in 1984, Mesezhnikov suggested a superdetailed zonation of the Ryazanian Horizon, which includes the following units (from the base upward): the *R. rjasanensis–Garniericeras subclypeiforme* Zone, the *R. rjasanensis–Hectoroceras kochi* Zone, the *R. rjasanensis–Surites spasskensis* Zone with three ammonite levels, and the *Surites tzikwinianus* Zone. Two newly introduced lower subdivisions were of crucial importance. *Garniericeras* and *Craspedites* forms from the horizon base are characteristic elements of the upper Volgian fauna, and confirm, in Mesezhnikov's opinion, the idea of N.A. Bogoslowsky and A.P. Pavlov that any significant hiatus between the Volgian and Ryazanian beds is missing. The erosion at the base of the Ryazanian Horizon is beyond any doubt; however, similar fauna from boundary beds suggests an insignificant time span for the respective hiatus in sedimentation.

Using these arguments, Mesezhnikov lowered the Ryazanian Horizon base down to the base of the *occi-*

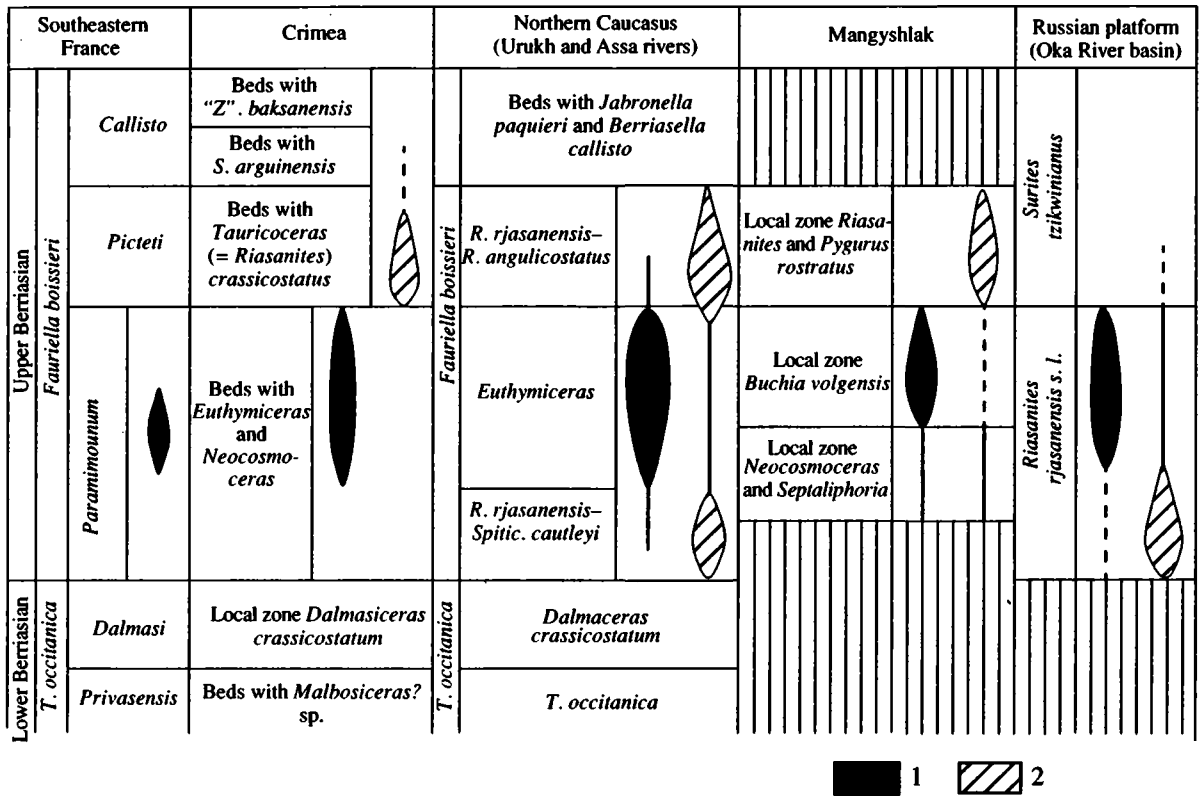

Fig. 2. Distribution ranges of *Euthymiceras* and *Riasanites* taxa in the southeastern France, Russia and CIS regions: 1, *Euthymiceras*; 2, *Riasanites*.

tanica Zone, and correlated the residual lower Berriasian interval with the *Craspedites nodiger* Zone of the upper Volgian Substage (*Zony yurskoi* ..., 1982; Mesezhnikov, 1984). To support this scheme and the correctness of correlation between basal levels of the *rjasanensis* Zone and *privasensis* Subzone, he cited the data of N.A. Bogoslowsky on the occurrence of *Berriasella* ex gr. *privasensis* together with *Riasanites* and *Euthymicera* species in the Oka River basin. However, Bogoslowsky (1895, p. 94) stressed that these species coexist in the upper interval of the Ryazanian Horizon, i.e., in the present *S. tzikwinianus* Zone. Moreover, the specimen of "*Hoplites* cf. *privasensis*", which was pictured by Bogoslowsky (now deposited in the TsNIGRI Museum, St. Petersburg), is more likely a *Riasanites* form with a peculiar morphology, than a species of the *Berriasella* genus. Furthermore, Mesezhnikov was of the opinion that correlation between the Berriasian and Ryazanian Horizon "can be clarified only after critical comparison of data on the Oka River basin, Crimea, Caucasus, and Mangyshlak" (Mesezhnikov *et al.*, 1979, p. 90). That is why he and his colleagues initiated their studies on the Berriasian Urukh section in the northern Caucasus, which provided materials very valuable with respect to the Boreal–Tethyan correlation. Unfortunately, the data obtained did not prove the preliminary conclusions of Mesezhnikov (1989).

The *Riasanites* and *Euthymiceras* are the main index taxa for the Ryazanian Horizon calibration against the Berriasian zonation in southern CIS regions and the European scale. As was shown, many stratigraphic schemes for the Berriasian in the Caucasus, Crimea, and Mangyshlak show the *Dalmasiceras* Beds marking the top of the *occitanica* Zone as overlain by the *Euthymiceras* Beds underlying the *Riasanites* (= *Tauricoceras*) interval. Another distribution pattern is characteristic of the Ryazanian Horizon: the *Riasanites* species are dominant in its lower part, whereas higher the *Euthymiceras* genus is abundant (Bogoslowsky, 1895; Mesezhnikov *et al.*, 1979). According to Mesezhnikov, this discrepancy points to different stratigraphic ranges of these taxa in different regions, and their correlation significance appears to be doubtful. Actually, the *Riasanites* distribution ranges in the Crimea, Caucasus, Mangyshlak, and Russian platform are distinctly incompatible (Fig. 2). In the northern Caucasus, this range is greater than elsewhere, and spans the main part of the *boissieri* Zone, presumably with the exception of the *callisto* Subzone. It shows here two abundance peaks of respective ammonites: the "lower" *Riasanites* level corresponding to the lower portion of the *paramimounum* Subzone, and the "upper" one likely comparable with the *picteti*

Subzone. Both peaks are separated by the prosperity interval of *Euthymiceras* forms.

Indubitably, the northern Caucasus was the radiation center of these peculiar berriasellids. The first migration wave of the *Riasanites* genus in the *paramimounum* time (the lower level) was oriented northward, when these ammonites penetrated, as suggested by Sasonova (1977), into the central Russian platform and central Poland through the Caspian depression. Initially rare *Euthymiceras* forms apparently migrated together with them. Sedimentation in Mangyshlak ceased at that time, and in the Crimea the lower *Riasanites* level is still undetermined. The peak migration was likely in the middle and end of the *paramimounum* time (the *Euthymiceras euthymi* Subzone), when the *Euthymiceras* genus colonized the Mangyshlak in addition to the Russian platform and Poland, and penetrated into southeastern France through the Crimea, Bulgaria, and Algeria. Finally, the upper *Riasanites* level corresponding to the prosperity time of respective ammonites is well-known in the Crimea and Mangyshlak, but they nearly ceased to migrate northward because of certain structural transformations.

Thus, the detailed correlation between the Ryazanian Horizon and Berriasian zonation in the southern CIS regions was hindered due to the inadequate interpretation of *Riasanites* distribution ranges. The distinguished lower *Riasanites* level was used in our scheme as a basis to correlate the *Riasanites rjasanensis* s. l. Zone of the Russian platform with the *paramimounum* Subzone, and the *Surites tzikwinianus* Zone with the *picteti* and *callisto* subzones of the stratotype *boissieri* Zone. Similar correlation schemes of subzones were suggested by Hoedemaeker (1987) and Bogdanova (*Zony melovoi* ..., 1989).

Nonetheless, data on fauna similarity and the insignificant hiatus between the upper Volgian and basal Ryazanian beds are valid, and dictate the necessity to raise the upper Volgian boundary up to the base of the Ryazanian Horizon. Mesezhnikov (1984, 1989) considered this operation to be poorly grounded, but not impossible. As we showed at the beginning, the ecotone faunas from the Russian Far East and North America include the middle Volgian buchiids associated with upper Tithonian ammonites (*Durangites, Kossmatia*) and the upper Volgian bivalves coupled with lower Berriasian *Pseudosubplanites, Berriasella, Dalmasiceras, Spiticeras, Proniceras*, and *Substeuroceras* genera. At present, we do not have reliable data to prove the synchronism of basal Berriasian and upper Volgian levels, although they seem to be very close. Hence, we correlate them with certain reservations.

Our correlation scheme is in good agreement with the results of Kutek and Zeiss (1974, 1975, 1988), who correlated Volgian and Berriasian deposits on the basis of mixed ammonite fauna in Poland, where the Volgian *Ilowayskya pseudoscythica* and *Zaraiskites scythica* zones yielded these index species in association with Tithonian *Isterites* and *Pseudovirgatites* species. As a reference section for the Mediterranean Tithonian, they used that of the Neuburg Formation in France (Barthel, 1975). Based on the distribution of *Isterites, Lemencia*, and some other ammonite taxa in the Neuburg section, and using data on Tithonian biostratigraphy in Spain (Enay and Geyssant, 1975; Oloriz and Tavera, 1982; Tavera, 1985) and Italy (Cecca *et al.*, 1986), Kutek and Zeiss correlated the *Isterites palmatus* level with the upper *Microcanthoceras ponti* Zone of the middle Tithonian, and further with the uppermost part of the terminal *pseudoscythica* Zone of the lower Volgian Substage. The parallel correlation for central European regions, such as southern Poland, Austria, Hungary, Slovakia, and Bulgaria, provided the same results being based on the distribution of *Pseudovirgatites* forms (Zeiss, 1977; Kutek and Zeiss, 1988; Zeiss and Bachmayer, 1989). In Poland, the upper part of the *pseudoscythica* Zone yielded early species of this genus, although this is most typical of the upper Tithonian, because the *Pseudovirgatites scruposus* Zone of Klentnice Beds (Austria) is well correlated with the *Simplisphinctes* Zone of Spain. Mesezhnikov (*Zony yurskoi* ..., 1982; *Zony melovoi* ..., 1989) and Jeletzky (1984, 1989), who argued against this correlation scheme, suggested their personal schemes and took *Zaraiskites* and similar forms for the true representatives of the genus. However, Kutek and Zeiss (1988) considered them to be homeomorphs of *Isterites* species, and this opinion is shared by such authorities in the European Tithonian such as Enay and Cecca.

All these data mean that the lower–middle Volgian boundary is the equivalent of the boundary separating the middle and upper Tithonian in central Europe (or its lower and upper substages in southern European regions). Subsequently, the middle and upper Volgian horizons should correspond to the upper Tithonian and lower Berriasian substages respectively (Zeiss, 1979, 1986), and we independently arrived at the same conclusion by analyzing stratigraphy and fauna of Jurassic–Cretaceous boundary beds in Russia and adjacent countries.

For the Russian platform, a similar correlation scheme (but with a different stage scale) was suggested by Sasonova and Sasonov (1983, 1984). They considered the Ryazanian Horizon as a stage synonymous with the range corresponding to the *boissieri* Zone, and attributed the upper Volgian Substage (the equivalent of the lower Berriasian *grandis* s.s. and *occitanica* zones) to the Kashpurian, and the lower–middle Volgian interval to the Gorodishchian regional stages. The upper portion of the latter was correlated with the *Berriasella jacobi* Zone considered now as a unit of the lower Berriasian (*jacobi/grandis* Zone). Unfortunately, the last idea and the correlation between the Kashpurian Stage and lower Berriasian were proved unsatisfactory, and specialists did not acknowledge this scheme.

NORTHERN SIBERIA

In contrast to the Russian platform, the Siberian sequence of Jurassic–Cretaceous boundary beds (Volgian Stage and Boreal Berriasian) is continuous. In the Boreal region of Russia, it was traditional to distinguish the Berriasian Stage at the base of the Cretaceous System as identical or nearly similar in range to this unit of the Mediterranean scale (Casey *et al.*, 1977; Sachs *et al.*, 1979; Shul'gina, 1984, 1985; *Zony melovoi* ..., 1989). The term *Boreal Berriasian* is recent and was proposed to stress a problematic correlation between the units in both provinces.

The continuous sections rich in ammonites are favorable for detailed stratigraphic studies and the correlation of respective deposits in Siberia and other Boreal regions. However, the Siberian scale cannot be directly calibrated according to the Mediterranean standard, although up to the present day, some southern elements among the upper Volgian ammonites of the region (*Virgatosphinctes* spp., *Aulacosphinctes*, and *Berriasella* now identified as *Lemencia*) have been considered as paleontological indicators suitable for this correlation. In Tethyan regions, these forms are typical of Tithonian deposits and do not cross the Tithonian–Berriasian boundary. This argument was decisive, when the upper Volgian Substage of Siberia was accepted as equivalent to the upper Tithonian, and the synchronism of Jurassic–Cretaceous boundary in both zoochores was taken for granted. Accordingly, the northern and southern Berriasian stages should be acknowledged as identical on account of the unbroken sedimentary sequences.

However, Zeiss (1968) and Casey (1973) doubted the *Virgatosphinctes* affinity of Siberian ammonites. As Zeiss suggested, the respective fossils represent a specific group probably younger and unknown among the Tithonian fauna. Sasonova and Sasonov (1979) distinguished the new genus *Praechetaites* comprising *Virgatosphinctes exoticus* Schulg. from Siberia as the type species, and suggested a Lower Cretaceous age for it. In our study on *Virgatosphinctes* from the southern Primor'e, we also came to the conclusion that Siberian forms are incompatible with true representatives of the genus. Moreover, according to available data on age constraints of *Virgatosphinctes*, the genus distribution is limited by the lower–middle Tithonian interval, and likewise the *Lemencia* genus does not cross the middle–upper Tithonian boundary in central and southern Europe.

Thus, we may correlate the Jurassic–Cretaceous boundary beds of Siberia with their European scale only by using data on regions with mixed Boreal–Tethyan faunas, such as the Russian platform, the Far East, and North America. The upper Volgian deposits of Siberia and the Russian platform can be correlated very well: (1) the *Craspedites okensis* Zone of Siberia should correspond to the *Kachpurites fulgens–Craspedites subditus* Zone, as the latter yielded *C. okensis*; and (2) the *C. taimyrensis* Zone is correlative with the *C. nodiger* Zone of the Russian platform, because both contain common species (Shul'gina, 1984, 1985). The cited author also believes that the *Chetaites chetae* Zone of Siberia may be partly comparable with the *nodiger* Zone, but most likely this level in the Russian platform is marked by a break in sedimentation (the pre-Ryazanian erosional event).

Correlation between the Ryazanian Horizon and the Berriasian beds of Siberia is more problematic. *Hectoroceras kochi*, which is widespread in the Boreal Realm, was found in Ryazanian beds (Casey *et al.*, 1977). This encountered the detailed zonal correlation of Boreal Berriasian and Ryazanian deposits, but it still remains disputable. According to one viewpoint, the *H. kochi* Zone of Siberia corresponds to the *Riasanites rjasanensis–H. kochi* Zone in the Russian platform, and the Berriasian in Siberia is equivalent to the Ryazanian Horizon (Casey *et al.*, 1977; Shul'gina, 1984, 1985; Zeiss, 1986). As we showed above, this horizon spans the range of the *boissieri* Zone, and this range should also correspond to the Berriasian in Siberia. Likewise, the upper Volgian Substage in this region should be correlated with the lower Berriasian *jacobi/grandis* and *occitanica* zones. Another standpoint (Hoedemaeker, 1987) is based on the detailed biostratigraphy of the *H. kochi* Zone, and suggests that the entire zone of *Riasanites rjasanensis* s.l. corresponds only to the upper part of the former.

In both cases, we should bear in mind that the position of *Hectoroceras* species in the Ryazanian section is not very clear. Mesezhnikov and his colleagues (Mesezhnikov *et al.*, 1979; Mesezhnikov, 1984) reported that the genus occurs below *Surites* and *Borealites* forms in the lower part of this horizon. At the same time, according to A.G. Olfer'ev and E.Yu. Baraboshkin (personal communication), *H. kochi* was found in its middle part (middle beds after Bogoslowsky) in association with *Surites* and *Borealites* genera[3].

Thus, the proper answer to the correlation problem of the *Hectoroceras kochi* Zone is as yet unknown. According to available data, its range may be equal to the main part of the *Riasanites rjasanensis* s.l. Zone, i.e., to the similar interval of the *paramimounum* Subzone of the Mediterranean scale (Fig. 3). The *Surites analogus* Zone of Siberia should be also correlated (in its total or partial range) with this subzone, because its index and some other species occur as well in the upper portion of the *Riasanites rjasanensis* s.l. Zone (the *R. rjasanensis–S. spasskensis* Zone after Mesezhnikov), where *Euthymiceras* forms are at peak abundance. An upper part of the *Surites analogus* Zone and the *Bojarkia mesezhnikowi* Zone may correspond in total to the residual upper Berriasian interval of the Mediterranean scale (*picteti* and *callisto* subzones).

[3] This and related problems are considered in more detail in our working notes on biostratigraphic criteria of the Jurassic–Cretaceous boundary (Sei and Kalacheva, 1993a).

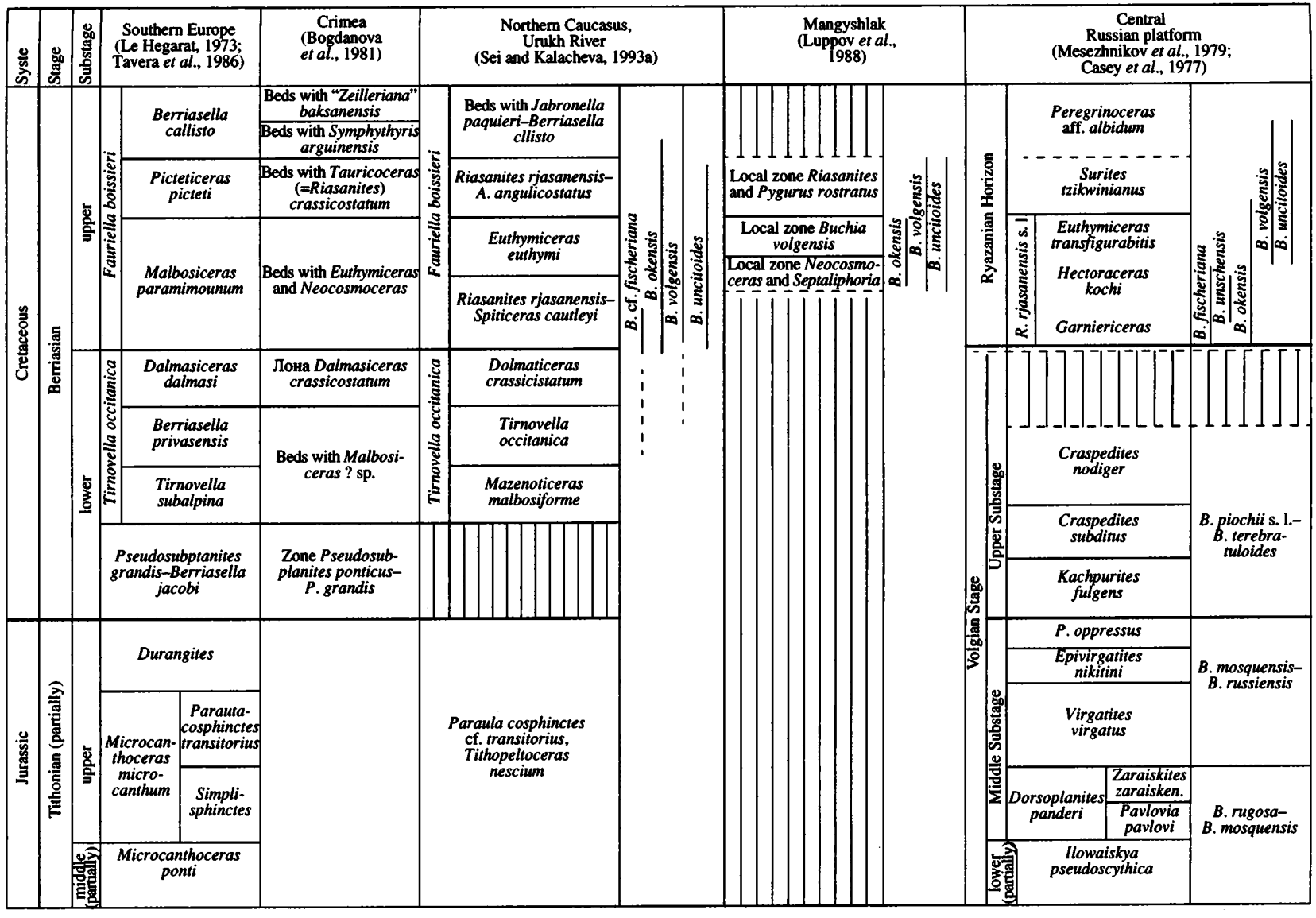

Fig. 3. Correlation scheme for Upper Jurassic–Lower Cretaceous deposits in southern Europe, Russia and other CIS countries, and America: 1, Jurassic–Cretaceous boundary now accepted for Boreal realm; 2, Jurassic–Cretaceous boundary in the Tethyan realm and the level suggested for Boreal regions.

Northern Siberia (Zakharov, 1981; Shul'gina, 1985)			Russian Far East (Sei and Kalacheva, 1993a, 1993b)		British Columbia (Jeletzky, 1984)	California (Oregon and Jones *et al.*, 1969)	Argentina (Leenza, 1980; Tavera *et al.*, 1986)	Mediterranean standard		
Boreal Berriasian	*Bojarkia mesezhnikowi*	*B. inflata*	*B. uncitoides*		*B. uncitoides*	*B. uncitoides*	*Spiticeras damesi*	*Fauriella boissieri*	Berriasian	Cretaceous
		B. tolmatschowi								
	Surites analogus	*B. jasikovi* (=*uncitoides*)			(*Neocosmoceras*)		*Argentiniceras noduliferum*			
	Hectoroceras kochi	*B. okensis*	*B. okensis*		*B. okensis* (*Argentiniceras*)	(*Neocosmoceras*)				
	Chetaites sibiricus	*B. unschensis*	*B. piochii s. l. – B. terebratuloides*		*B. terebratuloides* (*Parodontoceras, Substeuroceras*)	*B.* aff. *okensis* (*Parodontoceras, Substeuroceras, Proniceras*?)	*Substeuroceras koeneni*	*Tirnovella occitanica*		
Upper Volgian Substage	*Chetaites chetae*									
	Craspedites taimyrensis				*B. fischeriana*	*B. fischeriana* (*Spiticeras, Parodontoceras, Autacosphinctes*?)				
	Craspedites okensis	*B. obliqua* (=*terebratuloides*)		(*P.* cf. *grandis, B.* ex gr. *jacobi, Dalmasiceras sp.*)	*B. piochii* (*Proniceras*)			*B. jacobi–P. grandis*		
Middle Volgian Substage	*Epivirgatites variabilis*	*B. taimyrensis* (=*fischeriana*)	*B. russiensis–B. fischeriana* (*Durangites*)		*B.* cf. *blanfordiana*	*B. russiensis* (*Durangites, Kossmatia*)	*Corongoceras alternans*	*Durangites*	Tithonian (partially)	Jurassic
	Taimyrosphinctes excentricus		*B. mosquensis–B. russiensis*		*B. russiensis* s. l.			*P. transitorius*		
	Dorsoplanites maximum	*B. russiensis*								
	Dorsoplanites ilovaiskii	*B. rigosa–B. mosquensis*	*B. rigosa–B. mosquensis*		*B. mosquensis*		*Windhauseniceras internispinosum*	*Simplisphinctes*		
	Pavlovia iatriensis									
lower (partially)	*Pectinatites pectinatus*						*Aulacosph. proximus*	*M. ponti*		

1

2

Fig. 3. (Contd.)

At present, it is difficult to correctly evaluate the position of the *Chetaites sibiricus* Zone lowermost in the Boreal Berriasian. In European scales, it may span the boundary interval between the lower and upper Berriasian.

CONCLUSION

The performed correlation analysis involved stratotypes of southeastern France, Subtethyan localities in the Caucasus and Crimea, Subboreal sequences of the Russian platform and central Poland, Boreal sections of Siberia, and ecotones of the Russian Far East and North America, which can be interrelated with the stratotypes directly, or through Argentinean sections. The results show that Jurassic–Cretaceous boundary stages in Tethyan and Boreal realms are incompatible in their ranges. Our correlations suggest that lower and middle Volgian substages correspond to the standard Tithonian, while the Mediterranean Berriasian spans the upper Volgian Substage and Ryazanian Horizon, the latter approximately equivalent to the Boreal Berriasian (Fig. 3). Accordingly, the accepted position of the Jurassic–Cretaceous boundary should be revised in the Boreal Realm, or perhaps even across the globe.

The significance of Boreal–Tethyan correlations was in question for a long time. The idea to place this boundary at a level that can be reliably depicted in different zoochores was discussed many times. All meaningful levels between the Tithonian top and Valanginian base were inspected from this viewpoint. Following the priority principle, the International Working Group on the Jurassic–Cretaceous Boundary accepted the base of the *jacobi/grandis* Zone as most appropriate to discriminate these two systems. According to that decision, the preferred boundary can be replaced for any other, though close level, only if it will be impossible to trace the former in the Boreal realm (Remane, 1990).

For a long time, the high status of the *jacobi/grandis* Zone base raised doubts, because the underlying *Durangites* Zone was poorly studied. However, Spanish investigators (Oloriz and Tavera, 1982; Tavera, 1985; Tavera *et al.*, 1986) clearly demonstrated that the boundary in question is very distinct in its paleontological aspect. According to Oloriz and Tavera (1982, p. 154), "the change in ammonite fauna, taking place at that boundary, is considerable. The great explosion of Berriasellinae and the decline of Perisphinctidae and Himalayitinae indicate the type of ammonite spectrum that will be typical of the Berriasian. *Protacanthodiscus* is the only genus passing into the *jacobi* Zone." As we showed, this level in the Boreal regions most closely corresponds to the boundary between the middle and upper Volgian substages. The base of the upper substage is as clear as the Tithonian–Berriasian boundary. It marks the prosperity time of Craspeditidae with subfamilies Craspeditinae and Garniericeratinae, and also the decline of Virgatitinae and Dorsoplanitinae, which were very abundant in older Volgian beds (Shul'gina, 1985). The middle–upper Volgian boundary is similarly evident in *Buchia* fauna distribution patterns. It corresponds to the extinction of the *B. mosquensis* group, and to the decline of *B. russiensis;* only long-lived *B. fischeriana* and comparable *B. trigonoides* crossed this level. The base of the upper substage is also the plane, above which we can see nearly the concurrent appearance of *B. piochii* s.l., *B. terebratuloides*, and *B. unschensis*; the last two taxa are of Cretaceous morphology (Jones *et al.*, 1969; Jeletzky, 1984; Zakharov, 1986; Sei and Kalacheva, 1993b).

Among other levels, the most popular candidate for the Jurassic–Cretaceous boundary is the base of the *occitanica* Zone discussed by Mesezhnikov (1989), Zakharov (1986), and Hoedemaeker (1987). In their speculations, Mesezhnikov and Zakharov incorrectly assumed that basal levels of the *occitanica* Zone and Ryazanian Horizon are close or concurrent. Hoedemaeker elaborated this version in many details, however his Boreal–Tethyan correlations on the basis of *Buchia* acme zones in American and distant Siberian sections seem poorly grounded and not always correct. We believe that the idea to combine basal levels of the *occitanica* and *Chetaites sibiricus* zones is not very convincing, and attempts to trace this level in American sections were not impressive. At the same time, the base of the *occitanica* Zone in southeastern France is marked only by species combination and is therefore less distinct, in the paleontological sense, than the base of the *jacobi/grandis* zone.

Another, more reliable proposition is to accept the base of the *boissieri* Zone as indicating the Jurassic–Cretaceous boundary. It apparently coincides with the Ryazanian Horizon base in subboreal regions, and with basal levels of the *Hectoroceras kochi* and *Buchia okensis* zones in the Boreal realm. This version opens the door for Boreal–Tethyan correlations, but means a considerable decrease in the now accepted Berriasian range or even its replacement for the Ryazanian Stage (Zeiss, 1979; Remane, 1990). Besides, the basal *boissieri* level in the Mediterranean scale, like that of the *occitanica* Zone, is less evident than the base of the *grandis* s.l. Zone.

In opinion of J. Wiedmann, N.G. Khimshiashvili, N.I. Shul'gina, P.F. Rawson, and some other investigators, the Valanginian base also has a high correlation potential for being accepted as the Jurassic–Cretaceous boundary. However, this idea also meets with some difficulties (Zeiss, 1979; *Zony melovoi* ..., 1989). For instance, Company (1987) reported that among ammonite faunas from the Berriasian–Valanginian boundary beds of southeastern Spain (Mediterranean province), several typically Berriasian taxa, such as *Berriasella* s. str., *Tirnovella, Fauriella*, and *Jabronella*, pass into the basal Valanginian *otopeta* Zone. The characteristic Valanginian ammonite spectrum appears here only in the *pertransiens* Zone. Similar distribution patterns also seem to be typical of Boreal regions, where the Berria-

sian–Valanginian boundary is marked solely by changes in ammonite genera (Shul'gina, 1985).

Thus we believe that the base of the *jacobi/grandis* Zone, and, respectively, the middle–upper Volgian boundary in Boreal regions, are of the highest correlation potential among all stratigraphic levels in question, and represent the best level for the Jurassic–Cretaceous boundary. This raises a series of problems concerning the nomenclature of Jurassic–Cretaceous boundary stages in the Boreal scale. The situation with the Lower Cretaceous is not very complex, because the summary range of the upper Volgian Substage and Ryazanian Horizon, i.e., the Boreal Berriasian, is nearly equivalent to the Berriasian *sensu gallico*, and the term itself is not new for Boreal stratigraphy, although its current understanding suggests a somewhat different stratigraphic range.

By the assumed level of the Jurassic–Cretaceous boundary, the terminal stage of the Boreal Jurassic would include the lower and middle Volgian substages, and would be apparently equivalent to the Tithonian. This indicates the possibility of identifying the Tithonian deposits in the Boreal Realm and of achieving the global unification of Jurassic–Cretaceous boundary stages.

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The Continental Cretaceous Deposits in Northeastern Asia and Alaska: A Comparison of Floras and Paleoclimates

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Abstract—Flora from the upper beds of the Lower Killik and Upper Killik tongues and the Niakogon Tongue in northern Alaska is similar to the Grebenka flora of the Anadyr'–Koryak subregion (AKSR). Flora of the Tuluvak Tongue in northern Alaska is an age analog (Coniacian) of the Kaivayam flora in the AKSR; there is a significant difference between them, however, because the Tuluvak flora contains a considerable amount of the entire-leaved angiosperms. Floras of the Kogosukruk Tongue in northern Alaska and the Koryak and Gornaya Rechka floras in the AKSR are not exact age analogs, since they could have coexisted only during a certain interval of the Maastrichtian. The late Maastrichtian Koryak flora of the AKSR resembles the Alaskan Paleocene flora of the Sagwon section. The evidence derived from the AKSR Cretaceous flora allows reconstruction of climatic variations from the late Albian–Cenomanian warming to considerable Turonian–Coniacian cooling (with the maximum in the Turonian) then followed by new warming in the Santonian–early Campanian and subsequent cooling at the end of the Cretaceous. Estimated mean annual temperatures in northern Alaska were approaching 10°C for the Cenomanian, probably 13°C for the Coniacian, 5°C for the late Campanian–Maastrichtian, and 6–7°C for the Paleocene. In this area, the most significant differences in the character of climatic changes occurred in the Coniacian, when relatively low temperatures were in the AKSR, whereas the temperature maximum was in northern Alaska. This discrepancy is probably a result of significant changes in paleogeography that took place before the Coniacian and included the formation of the Bering Land, the isolation of the Arctic Basin from the proto-Pacific, and the opening of the Western Interior Seaway across the North American continent. Apparently, there was an oceanic current transfering heat along the seaway from the equatorial area to the Arctic basin. The depleted taxonomic composition of Maastrichtian flora in northern Alaska, as compared to that of the Gornaya Rechka and Koryak floras, is perhaps the result of a considerable temperature gradient in the circumpolar region at the end of the Cretaceous. The similarity between floras of the AKSR and Sagwon sequence in northern Alaska characteristic of the Cretaceous–Paleogene boundary beds testifies against the hypothesis of a global ecological crisis at this boundary time.

Key words: *paleobotany, paleoclimatology, paleogeography, Cretaceous, correlation, northeastern Asia, northern Alaska, Arctic basin, proto-Pacific, Western Interior Seaway*

INTRODUCTION

The biostratigraphy of the nonmarine Cretaceous deposits of northeastern Russia and Alaska, which cover vast areas in these territories, is based mainly on the paleobotanical data. Often, however, the comparison of fossil floras from these beds is difficult owing to the almost complete absence of common species and genera in respective assemblages.

The great differences between fossil floras that grew in the neighboring areas of a rather restricted region is attributed to the fact that these floras often imply very different, though adjacent, landscapes (Fig. 1) (Herman, 1993; Belyi, 1994). In the Late Cretaceous, part of the present territory of northeastern Russia and Alaska was covered by sea basins with coastal lowlands and well-developed river systems, floodplains, lakes, and swamps. At certain periods, the lowlands were flooded by a shallow sea with numerous islands. Further inland, there were uplands and plains where continental deposits accumulated in isolated depressions. The dissected highland of the Okhotsk–Chukotka volcanogenic belt stretched along the eastern margin of the Asiatic continent.

Floras of these diverse landscapes were, of course, very different. Since the Late Albian, in all the Late Cretaceous floras of northeastern Asia and Alaska that grew on the coastal plains, the angiosperms were dominant in abundance and diversity (Herman, 1993). During the Late Cretaceous, in continental depressions and on elevations of the volcanic upland, two kinds of floras coexisted: the "cenophytic" floras with predominance of the advanced angiosperms, which were more or less comparable with those on the coastal plain, and the "mesophytic" (Chauna and Arkagala) floras mainly composed of ferns and gymnosperms; the latter were so poor in angiosperms that often in their burials a single imprint of an angiosperm plant occurs together with dozens and even hundreds of fossil conifers.

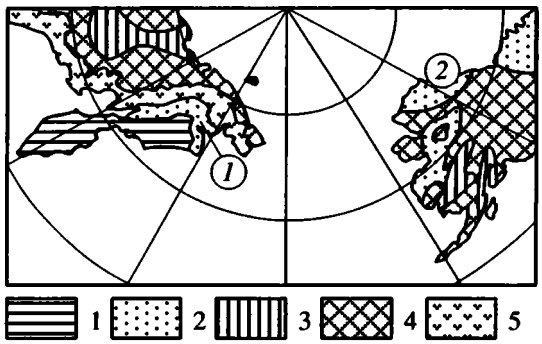

Fig. 1. Paleogeography of North Pacific Region in the Late Cretaceous (Belyi, 1994, with modifications; reconstructed position of continents at the Middle Cretaceous, after Smith *et al.*, 1981), and fossil flora localities in Alaska and the AKSR: 1, area of stable marine regime; 2, coastal plains and shallow shelfs; 3, non-flooded lowlands; 4, uplands and mountains; 5, Okhotsk–Chukotka volcanic belt; encircled numbers indicate distribution areas of plant-bearing beds in (*1*) the Anadyr'–Koryak Subregion and (*2*) northern Alaska.

The great difference in floras is probably caused by the ability of early angiosperms to colonize the disordered habitats of coastal lowlands with well developed river systems (Retallack and Dilcher, 1981, 1986). The vegetation of such lowlands was periodically destroyed by sea transgressions and, on a local scale, by migrations of the meandering river channels. These environments were favorable for the colonization of disordered habitats by the early angiosperms. Apparently, angiosperms gradually penetrated along the river valleys into the continent covered at that time by dominant coniferous–pteridophytic vegetation. The coexistence of different flora types in these regions at the Late Cretaceous presumably reflects a gradual invasion of advanced angiosperms into the continental interiors.

The objective of this paper is to compare the fossil floras of northeastern Asia and northern Alaska, because they developed under analogous conditions on coastal plains (Fig. 1) and, consequently, were similar. The age of the Cretaceous continental plant-bearing deposits of the AKSR and northern Alaska can be substantiated owing to their alternation with marine deposits. The continental beds often contain numerous and diverse plant remains, such as imprints of leaves and shoots, fossil woods, and remnants of fruits and seeds.

The intensive study of Cretaceous floras was started in the first half of this century. At that time, the most important results were obtained by A.N. Krishtofovich and A. Hollick. Study was continued further by V.A. Vakhrameev, T.N. Baikovskaya, L.Yu. Budantsev, N.D. Vasilevskaya, V.A. Krasilov, E.L. Lebedev, V.A. Samylina, G.G. Filippova, Ch.J. Smiley, and others, whose efforts have greatly contributed to our understanding of these floras. Since the end of the 1970s–begining of the 1980s, we also have been conducting research on the Cretaceous floras of northeastern Asia and Alaska. In recent years, we had the long-awaited opportunity for several joint comparative studies on collections of these fossil floras and for personal discussions of the results. This kind of cooperation is much more effective than the correlation of paleobotanical data on the basis of published materials. In the present paper, we describe the preliminary results of a comparative study of Late Cretaceous floras from the AKSR and northern Alaska (within the stratigraphic interval from the terminal Albian to the initial Paleogene).

The paleobotanical as well as lithological and geochronological data, are extensively applied in paleoclimatic reconstructions. We used the data on fossil floras of the AKSR and northern Alaska to outline the climatic changes that affected the composition and morphology of these floras (Spicer and Parrish, 1986, 1990; Parrish and Spicer, 1988; Herman, 1993). In this paper, we compare our results and discuss the possible reasons for similarities and differences between the paleotemperature curves plotted on the basis of paleobotanical data for the Late Cretaceous in northeastern Asia and northern Alaska.

MATERIAL AND METHODS OF STUDY

We and our colleagues L.B. Golovneva, E.L. Lebedev, G.G. Filippova, W. Lewis, M. Miller, J.T. Parrish, and others obtained many collections of plant remains (several thousands of specimens) from the Cretaceous and Paleogene deposits of Alaska and northeastern Asia, and used them as the material for our research. These collections are deposited at the Geological Institute of the Russian Academy of Sciences (RAS) in Moscow, the Botanical Institute, RAS, in St. Petersburg, the Sevvostok Geological Survey in Magadan, and the Open University in Milton Keynes, Great Britain.

The studied objects were mainly leaf imprints. The collections also contain fossil seeds, fruits, woods, and parts of shoots with leaves. They are often very well preserved and suitable for investigating the fine structural details of plant organs, but the phytoleims are, as a rule, lacking. The samples were photographed in oblique light with a small-format camera, and then black-and-white photographs of 20 × 25 cm were printed. In total, we used photographs of more than 3000 plants for our study. The leaf patterns were drawn right on the photographs with Indian ink, and the photographic image was bleached off with a water solution of potassium iodide and iodine, or with a water solution of potassium ferricyanide. The leaf drawings obtained were used in our analyses, and, via a scanner and computer, were recorded on CD-ROM for archives and further processing.

CORRELATION OF LATE CRETACEOUS FLORAS OF THE AKSR AND NORTHERN ALASKA

In the AKSR, which covers the territory of northern Kamchatka, the Koryak Highland, middle and lower reaches of the Anadyr' River, and the Pekul'nei Range (Fig. 1), the evolution of the Late Cretaceous (including the terminal Albian) flora (Fig. 2) is divided into the Grebenka stage (the end of Albian–? early Turonian), the Penzhina stage (Turonian, perhaps except its earliest part), the Kaivayam stage (Coniacian), the Barykov stage (Santonian–early (middle?) Campanian), the Gornaya Rechka stage (? late Campanian–middle Maastrichtian), and the Koryak stage (late Maastrichtian and, probably, Danian) (Herman, 1993; Golovneva, 1994). In northern Alaska, north of the Brooks Range (Fig. 1), fossil floras were recovered from the "tongues" (wedges of continental deposits in marine beds) studied at the following localities: the Upper and Lower Killik and Niakogon (late Albian–Cenomanian), Tuluvak (Coniacian), and Kogosukruk (late Campanian–Maastrichtian), and also in the Sagwon beds (Paleocene) of the Sagavanirktok Formation (Fig. 2) (Detterman *et al.*, 1963; Spicer and Parrish, 1990; Spicer *et al.*, 1992).

In northern Alaska, the first angiosperms appear only in the flora from the top of the Lower Killik Tongue (Spicer and Parrish, 1990). In the lower beds, the remains of angiosperms were not found. This part of the Alaskan sections probably corresponds to the lower part of the Ginter Formation in Ugol'naya Bay, which likewise lacks the remains of angiosperms (Herman, 1991). However, the last feature may be related to taphonomic peculiarities of these burials. A detailed comparison of floras at that level is impossible due to the impoverished flora composition in the lower beds of the Ginter Formation of the AKSR.

Flora from the upper beds of the Lower Killik Tongue and the Upper Killik and Niakogon tongues is similar in taxonomic composition to the Grebenka flora of the AKSR and the Arman' flora (Filippova, 1975; Herman and Shchepetov, 1994) of the Okhotsk–Chukotka volcanogenic belt north and northwest of the AKSR (Fig. 1); it can be, therefore, reliably correlated with these two floras (Fig. 2). This correlation concurrently indicates the same, or almost the same, ages of the Grebenka and Arman' floras. Even though they are not entirely identical, both display certain features of similarity with the Alaskan flora. Their age relations have been the subject of keen discussions for quite a long time (Samylina, 1988; Filippova and Abramova, 1993).

There are no analogues to the Penzhina flora of the AKSR in the north of Alaska, where Turonian deposits are represented by the Seabee Formation. The Coniacian flora of the Tuluvak Tongue in northern Alaska corresponds to that of the Kaivayam flora of the AKSR (Fig. 2). Despite a considerable difference between these floras, mainly manifested by the presence of entire-leaved angiosperms in the Tuluvak flora, they are well comparable on the basis of diverse platanoids and also *Paraprotophyllum, Trochodendroides, Magnoliaephyllum alternans* (Heer) Seward, *Dalembia pergamentii* Herman et Lebed., and some other taxa detected in their composition.

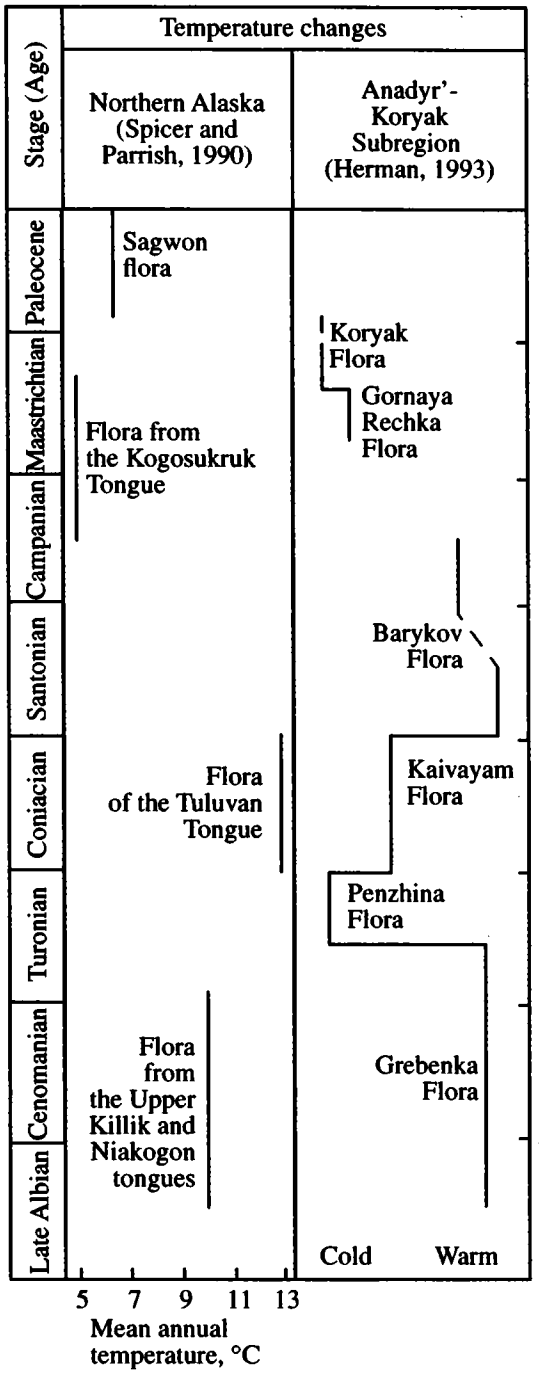

Fig. 2. Correlation of Cretaceous floras and reconstructed paleoclimatic changes in northern Alaska and the AKSR.

There are no analogues of the Barykov flora of the AKSR in northern Alaska, where Santonian and lower Campanian deposits are represented by the Schrader Bluff Formation containing the remains of marine mollusks.

Flora of the Kogosukruk Tongue in northern Alaska is not a complete age analogue of the Koryak and Gornaya Rechka flora of the AKSR, but they could have coexisted during a certain interval in the Maastrichtian (Fig. 2). This North Alaskan flora differs greatly from the AKSR floras in its extremely low taxonomic diversity: the pertinent assemblage consists of eight taxa, and only two of them are angiosperms. The plant assemblage of the late Maastrichtian Koryak flora of the AKSR is almost the same as the Paleocene Alaskan flora from the Sagwon sequence (Spicer *et al.*, 1994). Figure 2 shows the inferred correlation between the Cretaceous floras in the AKSR and those in northern Alaska.

LATE CRETACEOUS CLIMATIC CHANGES IN THE AKSR AND NORTHERN ALASKA

By studying the assemblages of Cretaceous floras in northern Alaska and the AKSR we were able to trace climatic changes that affected these regions (Spicer and Parrish, 1986, 1990; Parrish and Spicer, 1988; Herman, 1993, 1994). The Grebenka flora of the AKSR includes numerous plants that indicate a relatively warm climate: the cycadophytes *Nilssonia, Nilssoniocladus, Cycadites, Taeniopteris,* the conifer *Pagiophyllum* with lepidoid leaves, the entire-leaved angiosperms that constitute more than 50% of the flowering plants species, and others. The Penzhina and Kaivayam floras imply a cooler climate: the cycadophytes are rare or absent, the entire-leaved angiosperms are scarce, their burials contain mainly remains of large platanoid leaves with dentate margins. Evidently, these floras correspond to periods of cooling, the maximum of which falls, apparently, in the second half of the Turonian (the time of the Penzhina flora). The Barykov flora is essentially more thermophilic than the two previous floras: cycadophytes, entire-leaved angiosperms, and such dicotyledons as *Macclintockia* and *"Quercus" tchucotika* Abramova with small coriaceous leaves play a significant role in the composition of this flora. The Gornaya Rechka and Koryak floras show the progressive cooling of the climate toward the end of the Cretaceous. For instance, the Gornaya Rechka flora still includes cycadophytes, but among the angiosperms the plants with dentate leaf margins dominate (Golovneva, 1994), and further, the Koryak flora demonstrates a greater share of large-leaved angiosperms, no cycadophytes, and rare specimens of entire-leaved dicotyledons.

The studied changes in the Cretaceous flora of the AKSR allow us to reconstruct a relatively warm climate in the late Albian–Cenomanian and, possibly, in the early Turonian followed by a considerable cooling in the Turonian and Coniacian with the maximum in the Turonian, and then by warming in the Santonian and early Campanian, and by new cooling at the end of the Cretaceous. We did not calculate the possible mean annual temperatures using the content of entire-leaved angiosperms in the studied floras (Wolfe, 1979), but we can suppose that the mean annual temperatures varied from 6–8°C in the Turonian and at the end of the Maastrichtian to 11–13°C during the Cenomanian and Santonian–early Campanian optima.

To reconstruct the Cretaceous paleoclimate of northern Alaska on the basis of paleobotanical data we used quantitative and qualitative "climatic signals" (for more detail see Spicer and Parrish, 1990; Spicer *et al.*, 1992) to evaluate the mean annual temperatures for certain intervals within the Late Cretaceous period. The results were based on the well-known correlation between the percentage of the dicotyledonous species with entire leaves in the floras and the mean annual temperature (Wolfe, 1979). Among 67 leaved forms from the Upper Killik and Niakogon tongues, about a third have entire margins that approximately correspond to the mean annual temperature of 10°C. The Coniacian flora from two localites in the Tuluvak Tongue had a greater number of entire-margined angiosperms (10 of 25 taxa), which indicates a mean annual temperature of about 13°C. The low diversity of angiosperms in the flora from the Kogosukruk Tongue (late Campanian–Maastrichtian) does not allow temperature evalutions using the content of entire-leaved dicotyledons because samples appropriate for such an analysis should include no less than 20–25 taxa. However, taking into account the low diversity of this flora and comparing it with other Cretaceous floras of Alaska, we evaluate the mean annual temperature of this stage in a range from 2 to 6°C, probably 5°C. In the Early Tertiary flora from the Sagwon sequence, only one of 15 leaf forms of dicotyledons has an entire margin. This fact allows us to suggest the mean annual temperature of that period to be 6–7°C, although the insufficient sample size in this case may cause a considerable error.

Thus the approximate mean annual temperatures in northern Alaska are estimated to be 10° for the Cenomanian, 13° for the Coniacian, 5° for the late Campanian–Maastrichtian, and 6–7°C for the Paleocene.

DISCUSSION

Analyzing the obtained results we came to the following sequence of events (Fig. 2). The late Albian–Cenomanian climate of the AKSR and northern Alaska was rather warm, and perhaps even somewhat warmer in the AKSR owing to its more southern position and a certain isolation from the cold air masses of the Polar region and interior continental Asia thanks to the mountain ranges of the Okhotsk–Chukotka volcanic belt (Fig. 1).

In the Coniacian, a relative cooling in the AKSR was concurrent with the apparently extreme warming in northern Alaska, which is evident from a high abundance of entire-leaved angiosperms in the flora of the Tuluvak Tongue, which was situated at about 2°–3° to the north from the AKSR floras of the same age. Such a great contrast in the character of climatic changes obviously reflects essential paleogeographic reorganizations that occurred at the onset of the Coniacian stage and included the formation of the Bering Land, isolation of the Arctic basin from the proto-Pacific, and the opening of the Western Interior Seaway across the North American continent (Funnell, 1990). The seaway was apparently favorable for heat transfer through sea currents from the equatorial area to the Arctic basin (Fig. 3). These events could be responsible for a considerable warming of the Coniacian climate in northern Alaska located close to the junction of the Western Interior Seaway with the Arctic basin. According to the data by Hancock and Kauffman (1979), cited from Naidin *et al.* (1986), the Turonian and Coniacian were actually the periods of maximum transgression in the Western Interior Seaway. Finds of the Coniacian ammonites *Scaphites depressus* Reeside in the Kanguk Formation on the Axel Heiberg Island in Arctic Canada (Hills *et al.*, 1994) confirm that in the Coniacian, the Western Interior Seaway was connected with the Arctic basin. The narrowest part of the Seaway was no less than 500 km, while the widest parts were up to 2500 km wide (Hills *et al.*, 1994).

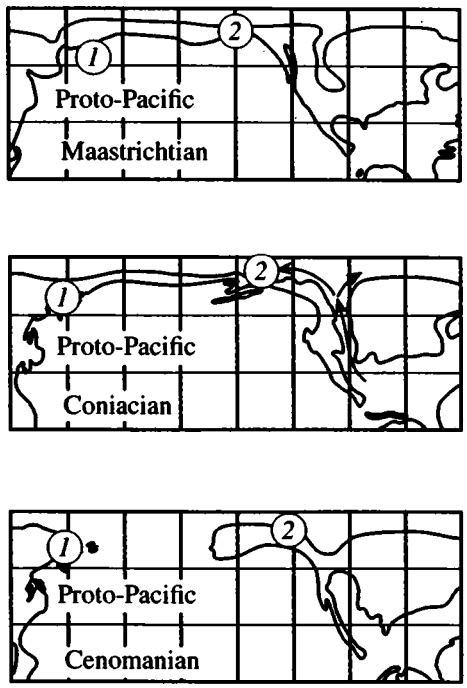

Fig. 3. Distribution of land areas and sea basins in the North Pacific Region and in western North America during the Late Cretaceous (after Funnell, 1990, with modifications); enciraled numbers indicate: (*1*) Anadyr'–Koryak Subregion and (*2*) northern Alaska; arrows show the assumed oceanic heat transfer through the Western Interior Seaway.

Moreover, in the Cretaceous the territory of northern Alaska was "open" toward the Arctic basin, which was perhaps sufficiently warm in the Coniacian, and isolated in the south by the Brooks Range, whereas the AKSR, as noted above, was open toward the proto-Pacific, which was, probably, cooler than the Arctic basin (Fig. 1). The existence of a rather abundant large-leaved Turonian flora in the Novaya Sibir' Island may also testify to relatively warm temperatures of waters in the Turonian–Coniacian Arctic basin (Sveshnikova and Budantsev, 1969; Herman, 1994). The morphological aspect of this flora evidently implies a temperate or cool-temperate climate, not colder than that of the Turonian time in the AKSR, though the locality of the Novaya Sibir' flora was on the paleolatitude of 82°N, whereas the paleolatitude of the Penzhina flora in the AKSR was near 72°N (Smith *et al.*, 1981).

There are substantially different appraisals of the role of oceanic heat transfer from the equatorial areas to polar regions. Some consider it to be the basic climate-forming factor, while others almost completely deny its significance (Nesov, 1992; Nesov and Yarkov, 1993; Covey and Barron, 1988). The latest data, obtained by Sellwood *et al.*, (1994), imply that this heat transfer could have been a more important element in the heat balance of the Earth in the Cretaceous than was assumed earlier. The paleobotanical data presented here testify in favor of a relatively warm Arctic basin in the middle of the Late Cretaceous and, probably, even during the entire Late Cretaceous excluding, perhaps, the terminal Maastrichtian. This warming in the Arctic basin was, apparently, the result of an intensive heat transfer by oceanic currents from the equatorial area through the Western Interior Seaway. The heat transferred by submeridional sea channels (the proto-Atlantic, the Western Interior Seaway, and/or the Turgai Strait) and consequent appearance of a sufficiently warm Arctic basin could have been responsible for the formation of glaciation-free (thermal type) climatic zoning in the Cretaceous, as was considered in detail by Chumakov (1995).

By the end of the Cretaceous, another cooling is recorded in both regions (Fig. 2), but the extremely limited floral diversity of the Kogosukruk Tongue indicates that it was much more cryophilic than the Gornaya Rechka and Koryak floras of about the same age. This can be attributed to the closing of the Western Interior Seaway and the resulting cooling of the Arctic basin at the end of the Maastrichtian, when northern Alaska became open toward this basin (Fig. 3), whereas the AKSR was isolated from it by the mountains of the volcanic belt. The more northern position of the Alaskan territory, which was, according to paleomagnetic reconstructions, near 85°N, is another relevant factor. The considerably depleted taxonomic composition of the Maastrichtian flora in northern Alaska, as compared

with that of the Gornaya Rechka and Koryak floras, probably shows a significant temperature gradient in the Polar regions at the end of the Cretaceous. Therefore, it seems rather difficult to explain occurrences of dinosaur remains far beyond the Polar Circle in the lower or middle Maastrichtian deposits of the AKSR and northern Alaska (Nesov and Golovneva, 1990; Brouwers *et al.*, 1987). Assuming the ability of dinosaurs to migrate, it is important to remember that faunal elements were not restricted in respect to their latitudinal distribution.

Nonetheless, the Paleocene flora from the Sagwon beds in northern Alaska is consistent in taxonomy and morphology with the late Maastrichtian Koryak flora of the AKSR. The latter, perhaps, was the source of many taxa for the Sagwon flora that migrated to northern Alaska from northeast Asia during the Paleocene warming (Spicer *et al.*, 1994). The similarity of two floras attributable to opposite sides of the Cretaceous–Paleogene boundary is evidence against the hypothesis of a global ecological crisis at this boundary (Spicer *et al.*, 1994). Possibly, it was the late Maastrichtian Koryak flora of the AKSR that gave rise to the Early Tertiary deciduous floras in the Arctic and adjacent regions (Herman and Spicer, 1995).

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Trends in the Evolution of Regional Faunas of Regular Archaeocyathids

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Abstract—Archaeocyathid fauna renewals are traced in eleven regions by dividing new genera of every chron into three groups: (1) immigrant genera, or colonizers; (2) aboriginal genera; and (3) genera, whose generation areas raise doubts. It is shown that dominant genera of the second group are important for detecting secondary centers of origin, while most abundant genera of the first group point to biotic invasions in the regions. The taxa prevalence in the third group is related to transgressions. Endemic genera mark the progressing specialization of taxa, but not the isolation of a region. As a rule, they appear in the centers of origin, and at the moments of peak diversity. Also analyzed are the growth, diversification, and regressive evolutionary stages of regional faunas.

Key words: *Archaeocyatha, Early Cambrian, regional fauna, endemism, centers of origin, specialization*

Our purpose in this study was to trace dynamics in the evolution of regional archaeocyathid faunas at the generic level. The class Archaeocyatha Regulares is a well-studied compact group, whose taxonomy has been elaborated in detail, and the group is suitable to be considered as a whole, but not by using its particular taxa. Current compilations on Archaeocyatha Regulares list 221 valid genera (Debrenne *et al.*, 1989, 1990), and we take into account data on their regional distribution from the latest publications (Perejon *et al.*, 1994; Eilicki and Debrenne, 1993; Debrenne, 1990; Mansy *et al.*, 1993).

The biogeography of archaeocyathids is analyzed in a series of works (Zhuravleva, 1968, 1981; Rozanov, 1976, 1980, 1984, 1992; Zhuravlev, 1984, 1986; Papeko *et al.*, 1992). Many of them are based on two principal ideas. The first suggests that archaeocyathids radiated within the equatorial belt between 30°N and 30°S. The second idea, though never formulated precisely, assumes that the geographic position of marine basins and their connections are of prime importance for the distribution and evolution of regional faunas. This simple and, at first glance, logical idea has still to be tested.

The analyzed survival curves and origin–extinction rates (Naimark *et al.*, 1995) proved the class of Archaeocyatha Regulares to display features of a real system, i.e., their evolution was under the control of some inner mechanisms in addition to external factors. Moreover, data on archaeocyathids, bryozoans, sea urchins, rugose corals, and cephalopods show that the distribution areas of their genera are related in a certain manner with the life time and number of genera in a higher taxon (Markov and Naimark, 1995). All these results suggest that the evolution of regional faunas is governed not only by regional paleogeographic factors, but also by development principles of the group as a whole; in other words, these faunas may reveal features of biotaxonomic systems (Markov and Naimark, 1994a, 1994b).

Archaeocyathid localities are known in eleven regions, such as the Siberian platform, the Altai–Sayan region, the Far East, the Mongolian region including Tuva and the Transbaikal area, northern Africa, western Europe, Australia, Central Asia including the southern Urals and Kazakhstan, China, and western and eastern North America. We cite these distribution regions in accordance with the classification of Debrenne *et al.* (1989, 1990). Actually, their discrimination seems to be sometimes artificial. For instance, the idea to unite the Urals, Kazakhstan, and Central Asia as a single region is simply the result of data scarcity for these areas, whereas it turned out to be convenient to discriminate northern Africa and western Europe by considering available materials. At the moment, it is well substantiated that these last regions were in the equatorial and subequatorial zones at the Early Cambrian time (Rozanov, 1980, 1984, 1992).

The successive colonization of all these regions is demonstrated in Figs. 1–6. To trace the evolution of regional faunas, we analyzed changes in genera composition within the aforementioned regions for every stage of the Lower Cambrian. Fourteen respective chrons were distinguished in total, arranged in the Tommotian and Atdabanian stages of four each, and in the Botomian and Toyonian stages of three each (*Yarusnoe* ..., 1984). We estimated the distribution of newly appearing genera in each of eleven regions. In each region, we divided new genera into three groups: (1) immigrants from other regions, i.e., genera whose earlier representatives are known in other regions;

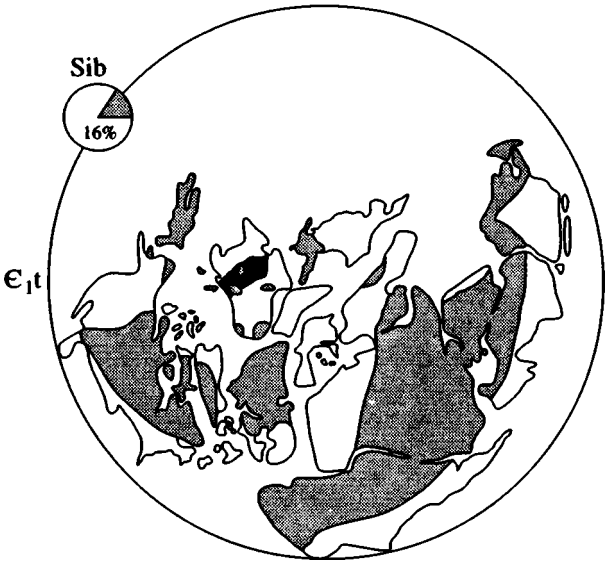

Fig. 1. Distribution areas of regular archaeocyathids (shown in black colour) during the Tommotian Stage. Here and in Figs. 2–6, paleogeography is after Rozanov (1984), and circular diagrams show percentages of endemic genera in the regions indicated by letter symbols: (Sib) the Siberian platform; (A–S) the Altai–Sayan region; (FE) the Far East, (Mon) Mongolia, Tuva, and Transbaikal area; (NAfr) northern Africa; (Aus) Australia; (NA) western North America; unfilled circles mean the absence of endemics at that time. Land regions shown by hatching.

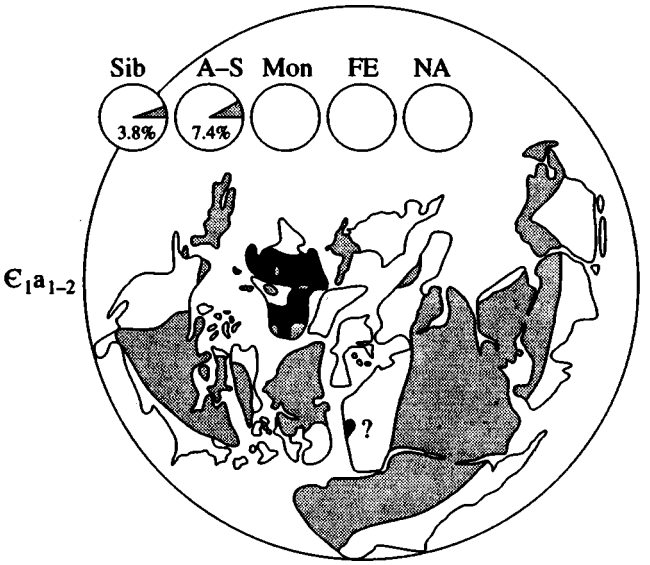

Fig. 2. Distribution areas of regular archaeocyathids during the early Atdabanian time.

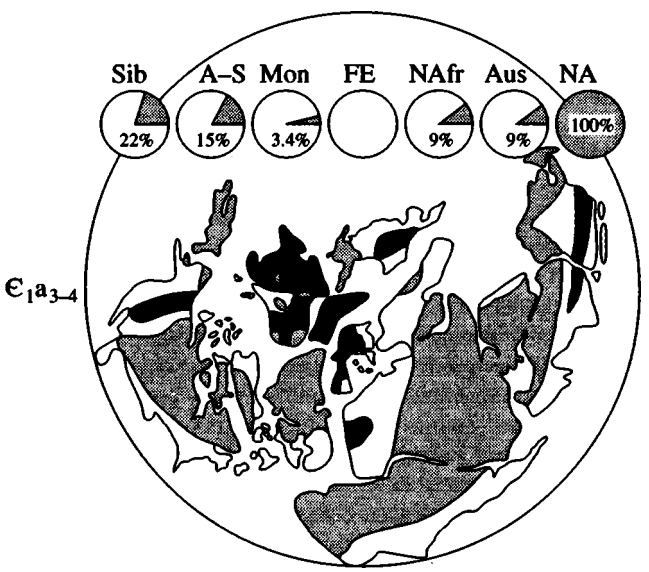

Fig. 3. Distribution areas of regular archaeocyathids during the late Atdabanian time.

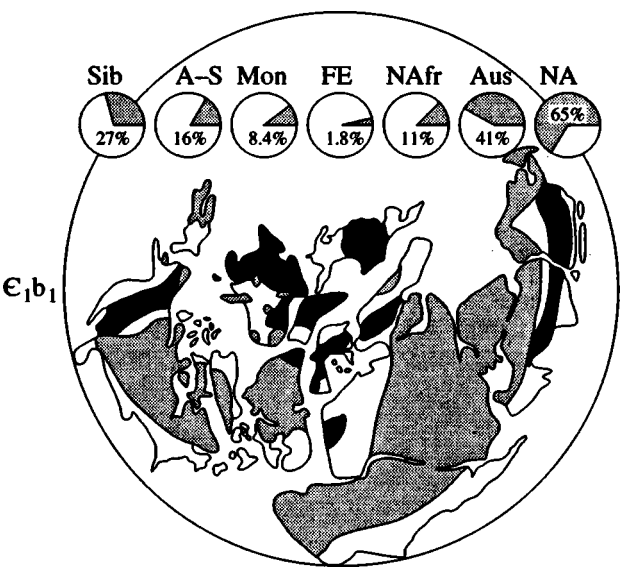

Fig. 4. Distribution areas of regular archaeocyathids during the early Botomian time.

(2) aboriginal genera, or the taxa first originated in the given region and apparently comprising endemic genera, which inhabited a single region during the whole history of their development; (3) genera displaying a wide radiation immediately after their origin in such a manner that it is impossible to establish their radiation centers. The results of our grouping are shown in Tables 1–10.

SIBERIAN PLATFORM

This region yields 71 genera of regular archaeocyathids.

The Siberian platform is the center of Archaeocyatha origin (Rozanov, 1980). The first genera appeared here at the beginning of the Tommotian Stage,

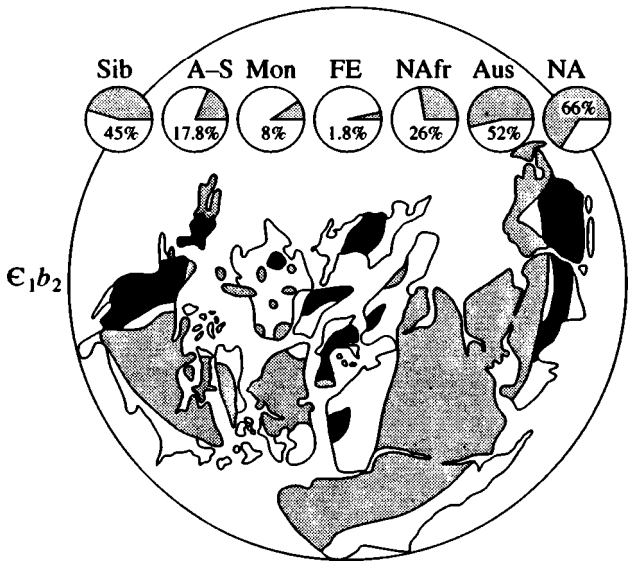

Fig. 5. Distribution areas of regular archaeocyathids during the middle Botomian time.

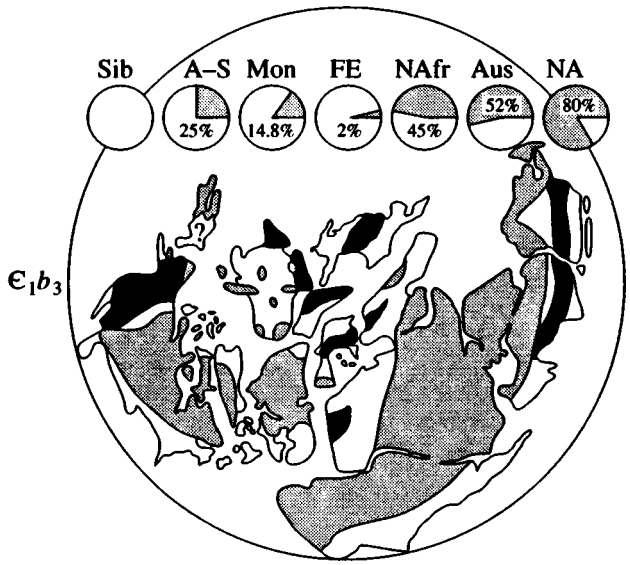

Fig. 6. Distribution areas of regular archaeocyathids during the late Botomian time.

and at the end of the stage their diversity corresponded to 12 genera. All these genera, with the exception of two very similar and endemic *Korshunovicyathus* and *Cryptoporocyathus*, radiated widely afterward. Naturally, all the Tommotian taxa fall into the second group.

In the first half of the early Atdabanian time (at_1), the diversity grew to 24 genera. All 12 new genera were native of the region (the second group); endemic taxa, however, did not appear. At this time, fauna in the Siberian platform was still isolated.

During the second interval of the early Atdabanian time (at_2), fauna in the platform included 26 genera. Among four new genera, one turned out to be endemic (*Fallocyathus*), and another widely radiated afterward (*Dokidocyathella*). Two new genera (*Plicocyathus* and *Coscinocyathus*), representing the third group, existed at the same time in the Altai–Sayan region and Mongolia.

Table 1. Dynamics of fauna renewal in the Siberian platform

Time	Total number of genera	New genera				
		total number	first group	second group	endemic genera in two groups	third group
tn	4	4	4	0	0	0
b_3	0	0	0	0	0	0
b_2	12	1	0	0	0	1
b_1	54	17	1	8	8	8
at_4	40	14	4	8	6	2
at_3	28	7	1	2	2	4
at_2	26	4	0	2	1	2
at_1	24	13	0	13	0	0
t	12	12	0	12	2	0

Note: Here and in tables 2–10, stage symbols are as follows: (t) Tommotian; (at_{1-4}) Atdabanian; (b_{1-3}) Botomian; (tn) Toyonian.

In the middle Atdabanian time (at_3), there were 28 Siberian genera, among which seven were new. Distribution of these taxa between our groupings is shown in the table. We should remark that the middle Atdabanian was the time of the first newcomers in Siberia. These were representatives of the *Gordonicyathus* genus, whose first occurrence level is recorded in the Altai–Sayan region. The third group included four genera at this time. Three of them simultaneously existed in the last region, two in northern Africa, and one in Mongolia.

The Siberian platform of the late Atdabanian time (at_4) accommodated 40 genera; 14 of them were previously unknown here. Two genera among the latter were native of the region, one concurrently appeared in Siberia and Altai–Sayan region, and one more in Siberia, Mongolia, and western Europe. Endemic genera grew remarkably in number up to six. Also striking was activated immigration from other regions: two genera are newcomers from the Altai–Sayan region, and we do not rule out immigration from northern Africa, because the genus *Tumulifungia* is likely an aboriginal taxon of this region.

The early Botomian (b_1) archaeocyathids were represented by 54 genera; 17 among them did not occur in the region before this time. In the third group, new genera were distributed in the following manner: two genera were in common with the Altai–Sayan region, two with Mongolia, two more with the Far East, and three genera showed their appearance almost everywhere soon after their origin. We also noted eight new endemic taxa. The early Botomian in the Siberian platform was

marked by the invasion of *Pretiosocyathus* forms, whose origin is related to the Altai–Sayan region.

The middle Botomian time (b_2) marks the decline in diversity of Siberian archaeocyathids: many of the 12 surviving genera occur only in basal middle Botomian beds. Endemics did not appear at that time, and one sole genus *Erbocyathus* was new in the region. This genus appeared at once in three regions, namely in the Siberian platform, the Altai–Sayan region, and Mongolia. In the late Botomian time, none of the genera survived in the Siberian platform.

The Toyonian Stage in Siberia is marked by four genera, two of which initially appeared in the Altai–Sayan area, and such taxon as *Kiwicyathus* probably immigrated from Australia. The fourth genus *Irinaecyathus* was not absolutely new to the region, because it was here prior to the middle Botomian time, then disappeared and came again, but already as an immigrant from another region. Endemic genera of Toyonian time are unknown. We do not consider here genus *Angaricyathus*, because its host rocks in original locality are of unclear age, and the known species from Morocco is of doubtful affinity with this genus.

Thus, all the data show that regional archaeocyathid fauna is conceived at the beginning of the Tommotian Stage. Genera of that time had a high potential for ecological adaptation, and therefore became widely dispersed afterward. In addition to these "general" genera, there were also endemic taxa. These were genera apparently adapted to the particular conditions of the Siberian platform, but less endurable than other Tommotian archaeocyathids. They became extinct in the terminal Tommotian, or incipient Atdabanian time.

The fauna renewal across the Tommotian–Atdabanian boundary was at the expense of aboriginal archaeocyathids. Later, it was promoted by biota migration from the Altai–Sayan region, and this was responsible for a growing similarity between faunas of Siberia and other regions. At the mid-Atdabanian time, taxa of the Siberian platform comprised forms common to the North African fauna. During the early Botomian, fauna diversity in the region was increasing and then approached the peak point. The higher abundance obtained at the expense of cosmopolitan genera of that time (third group) was enhanced by more active production of endemic taxa. An additional peculiarity of this interval is evident from the fact that no newly born genera of this time can be related in origin with the Siberian region.

The remarkable middle Botomian decline of archaeocyathid fauna in the Siberian platform is the result of considerable environmental changes in marine basins. Accordingly, fauna renewal almost ceased in the middle–late Botomian time. It became somewhat activated during the Toyonian Stage. The Toyonian archaeocyathids were immigrants from the Altai–Sayan region, and presumably from Australia. After the middle Botomian time, endemic genera are missing among Siberian archaeocyathids.

Table 2. Dynamics of fauna renewal in the Altai–Sayan region

Time	Total number of genera	New genera				
		total number	first group	second group	endemic genera in two groups	third group
tn	4	0	0	0	0	0
b_3	40	0	0	0	0	0
b_2	84	7	1	5	4	1
b_1	90	30	2	13	10	15
at_4	71	16	3	7	3	6
at_3	55	29	5	17	6	7
at_2	27	27	15	9	2	3

ALTAI–SAYAN REGION

The number of regular archaeocyathid genera, which have been recorded in this region, is 110 (Table 2), and their first representatives appeared here in the second half of the early Atdabanian time (at_2). Initial colonization of the region was rather vigorous, as already in the early Atdabanian there were 27 genera, more than anywhere else at that time. The dominant group (15 genera) consisted of newcomers, which clearly were inborn taxa of the Siberian platform. It is very likely and interesting that five of them first colonized the Far East, and only then settled in the Altai–Sayan Mountains. The second group included nine genera, and two of them were endemics. Three genera of the next group concurrently appeared in the Altai–Sayan, Mongolian, and Siberian regions.

In the middle Atdabanian time, the diversity was twice as high (55 genera); 29 among these taxa were new. The group of aboriginal genera (17) was dominant, and included six endemic taxa. The first group of distinct immigrants was reduced down to five genera. Two of them originated from northern Africa, and one is recorded in both the North African and Altai–Sayan regions, however in the former it seems to be more ancient. Within the interval in question, one genera of the third group (*Mennericyathus*) occurred in the Altai–Sayan Mountains, northern Africa, and the Siberian platform. Fauna of the region also retained some affinity with archaeocyathid assemblages of the Far East and Mongolia (Papeko *et al.*, 1993).

The late Atdabanian archaeocyathids in the Altai–Sayan region were represented by 71 genera. The diversity growth and fauna renewal were mainly characteristic of the second group (7 genera, among them three endemics). The third group included six genera, which were concurrently dispersed in the Far East, Mongolia, and the Siberian platform in addition to the Altai–Sayan region. Finally, three extra genera had their roots in other regions: one in northern Africa, and two in the Siberian platform. These Siberian taxa initially (at the middle Atdabanian time) migrated to the Far East, and

only afterward colonized the area in question. The same radiation scheme was characteristic of the early Atdabanian time.

The early Botomian fauna of the Altai–Sayan area incorporated 90 genera, including 30 new taxa. The third group was dominant and made up half of the new genera. Simultaneously abundant were aboriginal taxa mainly of the endemic type. The first group consisted of two genera only: *Zonacyathus* born in the Siberian platform in the late Atdabanian time, and *Favilynthus*, which appeared at the same time in the Far East.

The middle Botomian time with its 84 genera marks the initial diversity decrease and decelerated fauna renewal, as only seven new taxa appeared. The second group contributed much—exactly five genera, four of which were endemic. One genus *Sajanocyathus* was an immigrant of presumably Australian origin. The next *Erbocyathus* genus came into view concurrently in the Altai–Sayan region, the Siberian platform, and the Far East.

A prominent decline in diversity of regular archaeocyathids took place here in the late Botomian time, when only 40 genera survived, and the rest became extinct. Fauna renewal completely ceased. No new genera came into existence in the region at the Toyonian Stage, and only four taxa survived. Two of these taxa were widespread at the initial stages of their evolution: *Vologdinocyathus* apparently originated from Australia, and *Tegerocyathus* was of native descent. All endemics died out.

Thus, archaeocyathids of the Altai–Sayan region showed incipient development at the early Atdabanian time. During the first stage, they mostly consisted of immigrants from Siberia. In the middle–upper Atdabanian interval, the situation changed. This period is featured by decelerated immigration, and by the origin of native genera capable of invading vast territories. Those few genera that still migrated in the region, had their roots in northern Africa and Siberia. Cosmopolitan taxa attributable to our third group increased in number at the early Botomian time. It is well evident that expansion of this group, indicating a greater similarity between faunas from different regions and a lower degree of their isolation, was not related to the development of endemics. However, fauna renewal at the middle Botomian time was caused by the appearance of endemic taxa. In general, the Botomian fauna of the Altai–Sayan region was impoverished, as compared to the Atdabanian interval, in those genera that subsequently colonized other regions.

MONGOLIA, TUVA, AND THE TRANSBAIKAL REGION

As much as 80 genera are known from this composite territory (Table 3). As one can see from the table, the territory represented the habitat of immigrated taxa. During all the stages, immigrants were dominant among other genera appearing here for the first time. Before the middle Atdabanian time, newcomers were mostly native taxa of the Siberian platform, and afterward they had their roots in the Altai–Sayan region. That is to say, the Mongolian fauna was most common in character with archaeocyathids of the Altai–Sayan region, and genera shared by both areas were at maximum abundance in the early Botomian time. The second group of Mongolian taxa consisted solely of endemics. The latter appeared in the region in the middle Atdabanian–middle Botomian time. Their peak occurrence corresponds to the early Botomian, i.e., to the time when immigration was considerably accelerated, and similarity with other regions was most pronounced. In contrast, the diversity of the late Botomian regional fauna was extremely reduced, and only one new genus is recorded in its composition. It was the genus *Tegerocyathus*, an immigrant from the Altai–Sayan region, where it resided until the termination of the Toyonian Stage.

The mean distribution area, we estimated for genera newly coming in the region, was gradually decreasing. The local peak of this parameter was confined to the early Botomian time, although the sole late Botomian newcomer also occupied a considerable area in the region. As was shown before (Markov and Naimark, 1995; Naimark and Markov, 1995) the mean distribution area is apparently indicative of the specialization degree of a genus with respect to other taxa of the group. Therefore, the decreasing values of this parameter suggest the region to be successively invaded by more and more specialized genera. Against this background, the early Botomian time marks a somewhat accelerated input of nonspecialized, or "general" taxa. Judging from the distribution area index, the last newcomer, *Tegerocyathus*, apparently was of a low specialization rank.

THE FAR EAST

This region is known as accommodating 61 genera. Its colonization began in the early Atdabanian time (at_1). Like in the Mongolian region, fauna of the Far East was expanded by immigrants from other regions. Three successive endemic genera appeared here during the Botomian Stage. Before the middle Atdabanian time, the region was mainly colonized by inborn taxa of the Siberian platform, however taxa dominant afterward were native genera of the Altai–Sayan region. The genus *Inessocyathus*, whose representatives migrated to the Far East in the middle Atdabanian time, is known at the moment from older rocks only in northern Africa. Genera of the third group were added to the fauna in question in the late Atdabanian–early Botomian time. The diversity peak of archaeocyathid fauna in the Far East was at the middle Botomian time. The best similitude was characteristic of this and Siberian faunas, as both included 37 common genera, or 59% of their generic spectrum.

The mean distribution area for genera immigrated in the Far East gradually decreases with time, though with

local peaks confined to the middle Atdabanian and early Botomian times. This indicates that growing specialization of archaeocyathids in the Far East was complicated by the addition of more general taxa during the middle Atdabanian and lower Botomian intervals. As noted already, *Tegerocyathus*, the sole newcomer of the late Botomian time, was widely distributed. Like in Siberia and northern Africa, the archaeocyathid decline in the Far East started at the mid-Atdabanian time (at_3), while in the Mongolian and Altai–Sayan regions this happened during the late Atdabanian.

NORTHERN AFRICA

Fauna of this region, comprising 37 genera in total (Table 5) was formed in the first half of the Atdabanian Stage (at_2). In the early Atdabanian time, native taxa were insignificantly dominant. Four immigrants (*Nochoroicyathus, Sibirecyathus, Erismacoscinus*, and *Retecoscinus*) represented nonspecialized taxa from Siberia. They coexisted with the single endemic genus *Geniculicyathus*.

During the middle Atdabanian, the diversity of North African archaeocyathids was growing due to both immigrants from Siberia and taxa from the third group. Two of three taxa of the last group concurrently appeared in Siberia, and one in western Europe. The late Atdabanian genera of the region show an accelerated extinction rate and a cessation of fauna renewal.

The renewal was activated again in the early Botomian interval, and 11 new genera included two endemics, three taxa of the third group, and six immigrants. The newcomers took their origin in the Siberian platform (5 taxa) and Altai–Sayan region (1 taxon). The maximum fauna diversity is recorded in middle Botomian strata with all three groups participating almost equally. Immigrants (2 genera) were from the Altai–Sayan Mountains, although the region in question was usually open for taxa migrating from Siberia. The fauna renewal terminated in the region during the initial late Botomian time. No genera remained here in the Toyonian Stage.

AUSTRALIA

Colonization of Australia, accommodating 70 genera of regular archaeocyathids, was initiated in the late Atdabanian time (Table 6). The main part of twelve late Atdabanian taxa (10 genera) included immigrants from other regions, mostly from Siberia (6 genera). This chron also marks the origin of the Australian genus *Prethmophyllum*, which subsequently colonized basins of West European and Mongolian regions.

The considerably activated fauna renewal here was in the early Botomian time, when 43 genera, including 22 endemics, appeared in Australia for the first time. Eleven immigrants of this period included six native Siberian taxa and four newcomers from the Altai–Sayan region. The middle Botomian renewal was predominantly due to endemic genera, and this time marks the peak abundance of archaeocyathids in the region. Afterward, there was a quick decline of fauna diversity in Australia, and its renewal almost ceased. The sole new genus of this time (*Vologdinocyathus* from the third group) was able, however, to colonize four regions.

Australian archaeocyathids of the Toyonian Stage are represented by two genera only. The last new genus of the region is *Kiwicyathus*, which appeared at the beginning of the stage and subsequently colonized the Siberian platform.

The mean distribution areas, which we estimated for both new and all Australian genera, were decreasing from stage to stage, and showed an insignificant increase only at the beginning of the Toyonian time. This points to the progressively growing specialization of Australian fauna.

WESTERN EUROPE

Archaeocyathid fauna of western Europe includes 39 genera (Table 7). It was at its initial stage of evolution at the middle Atdabanian time (at_3). Throughout its

Table 3. Dynamics of fauna renewal in the Mongolian region

Time	Total number of genera	New genera				
		total number	first group	second group	endemic genera in two groups	third group
tn	4	0	0	0	0	0
b_3	28	1	1	0	0	0
b_2	61	2	0	1	1	1
b_1	72	34	18	6	6	10
at_4	43	19	13	1	1	5
at_3	24	16	12	1	1	3
at_2	8	8	6	0	0	2

Table 4. Dynamics of fauna renewal in the Far East

Time	Total number of genera	New genera				
		total number	first group	second group	endemic genera in two groups	third group
tn	2	0	0	0	0	0
b_3	26	2	1	1	1	0
b_2	47	8	7	1	1	0
b_1	44	22	16	1	1	5
at_4	27	9	5	1	0	3
at_3	20	10	10	0	0	0
at_2	10	4	3	0	0	1
at_1	6	6	5	1	0	0

Table 5. Dynamics of fauna renewal in northern Africa

Time	Total number of genera	New genera				
		total number	first group	second group	endemic genera in two groups	third group
tn	0	0	0	0	0	0
b_3	11	0	0	0	0	0
b_2	28	9	2	2	2	3
b_1	24	11	6	2	2	3
at_4	16	3	2	0	0	1
at_3	16	7	3	1	0	3
at_2	9	9	4	5	1	0

Table 6. Dynamics of fauna renewal in Australia

Time	Total number of genera	New genera				
		total number	first group	second group	endemic genera in two groups	third group
tn	2	1	0	1	0	0
b_3	46	1	0	0	0	1
b_2	63	13	3	10	10	0
b_1	56	43	11	23	22	9
at_4	13	12	10	2	1	0
at_3	1	1	1	0	0	0

Table 7. Dynamics of fauna renewal in western Europe

Time	Total number of genera	New genera				
		total number	first group	second group	endemic genera in two groups	third group
tn	0	0	0	0	0	0
b_3	9	0	0	0	0	0
b_2	22	2	0	0	0	2
b_1	26	15	5	4	4	6
at_4	22	9	6	0	0	3
at_3	13	13	10	1	1	2

history, this region was a refugium for immigrants, and its dominant genera were newcomers from other regions. Their first representatives were mostly taxa of Siberian origin (9 of 10 genera of the first group), which coexisted with one endemic genus. Two middle Atdabanian taxa of the third group (*Gordonifungia* and *Afiacyathus*) occurred, beyond western Europe, in the North African region. Nine new genera appeared in the European area at the late Atdabanian time. Dominant among them were immigrants (6 genera) including four taxa originating in the Siberian platform, one genus from the Altai–Sayan area, and another from northern Africa. The initial and rather quick extinction of archaeocyathids in the region was across the Atdabanian–Botomian boundary interval, when 11 genera died out. At the same time, this period is marked by the intense renewal of archaeocyathid taxa, as 15 new genera appeared in the early Botomian time.

The middle Botomian archaeocyathids include two new genera, whose forms are concurrently recorded in North African assemblages. Cessation of fauna renewal in western Europe was in the late Botomian time, when it first regressed, and then, at the end of the Botomian Stage, the last regular archaeocyathids became extinct.

The mean distribution areas of new genera in the region decreased during three successive chrons (at_3, at_4, and b_1). It is also noted that, in contrast to other regions, this parameter did not rise here in the early Botomian interval. Apparently, this was related to the fauna regress at the earlier stage due to some specific regional conditions.

NORTH AMERICA

In western North America (Cordillera and Koryak Upland), there were described 17 genera (Table 8). Archaeocyathid evolution was initiated here only at the end of the Atdabanian Stage. As one can see from the Table 8, the first taxa of the region were endemic without a single immigrant genus. This presents a problem concerning the roots of West American archaeocyathids. Data available at the moment do not provide any reasonable solution. In the early Botomian time, the region was marked by the appearance of 13 new genera, and more than half of them were endemic (7 taxa). One of the endemics, the genus *Robertiolynthus*, is the only representative of the order Monocyathida, and accordingly, the origin of the genus is unclear. The same can be stated with respect to the endemic genus *Acanthopygrus*, the sole taxon of the order Capsulocyathida in the area. At the same time, western North America was colonized as well by resistant cosmopolitan genera *Nochoroicyathus, Ajacicyathus*, and *Plicocyathus*. Among five immigrated genera, three were native taxa of the Siberian platform, one migrated from Australia, and one more originated presumably in the Altai–Sayan region.

The fauna renewal in western North America was completed in the middle Botomian interval. This was a period of fauna regress and taxa extinction. In our opinion, the early decline of archaeocyathids was in response to peculiar regional factors, like in western Europe.

In eastern North America, archaeocyathid assemblages are very impoverished and represented by three genera only. Two of them are immigrants (*Tegerocyathus* and *Subtilocyathus*), which appeared here in the

late Botomian time. Another genus, *Vologdinocyathus*, migrated to the region during the middle Toyonian Stage. Endemic genera are missing in this area.

CENTRAL ASIA, SOUTHERN URALS, AND KAZAKHSTAN

The total number of genera recovered in these areas is 20 (Table 9). The history of archaeocyathids begins here in the middle Atdabanian with the appearance of *Archaeolynthus* and *Nochoroicyathus* genera, whose first representatives in Siberia are recorded in Tommotian beds. The late Atdabanian assemblage from these Asian regions also includes another Siberian genus *Orbicyathus*.

At the beginning of the Botomian Stage, the area in question was colonized by 14 new genera: 12 of them were immigrants and two represented the third group. Endemics were absent. Colonizers included six native taxa of the Siberian platform, two genera from the Altai–Sayan region, two others from northern Africa (*Tumulifungia* and *Degeletticyathus*), and two additional taxa of unclear origin (*Sanarcocyathus* and *Mennericyathus*). Genera that colonized Asian areas during the middle–late Botomian time originated in the Altai–Sayan region. The total Asian territory under consideration was barren of endemics. The decline of archaeocyathids was triggered here at the middle Botomian Stage. In general, the territory represented an area of colonization, and its fauna was rather poor.

CHINA

The Chinese fauna of regular archaeocyathids was even less diverse than that in central Asia. It included 16 genera, and among three known orders of Archaeocyatha Regulares only one Ajacicyathida is recorded now in China. The first Chinese archaeocyathids were colonizers from the Siberian platform and migrated into the region during the late Atdabanian time. The moment of significant diversification at the generic level is recorded in the lower Botomian beds and was a result of migration from the Siberian platform, the Altai–Sayan region, northern Africa, and western Europe. The peak diversity in the middle Botomian time was mainly due to genera survival rather than fauna renewal. Endemics are unknown in this region (Table 10).

SECONDARY CENTERS OF ORIGIN OF REGULAR ARCHAEOCYATHIDS

Our procedure of subdividing all genera present in the regional record into colonizers, aborigines, and fast-radiating taxa affords an opportunity to analyze processes or fauna evolution, which were dominant in each region at one stage or another. The prevalence of immigrants (genera of the first group) indicates that a given region was in a state of colonization. Dominant

Table 8. Dynamics of fauna renewal in North America

Time	Total number of genera	New genera				
		total number	first group	second group	endemic genera in two groups	third group
b_3	5	0	0	0	0	0
b_2	15	0	0	0	0	0
b_1	17	13	5	7	7	1
at_4	4	4	0	4	4	0

Table 9. Dynamics of fauna renewal in the central Asian region

Time	Total number of genera	New genera				
		total number	first group	second group	endemic genera in two groups	third group
tn	2	0	0	0	0	0
b_3	8	2	1	0	0	1
b_2	18	2	1	0	0	1
b_1	17	14	12	0	0	2
at_4	3	1	1	0	0	0
at_3	2	2	2	0	0	0

genera of the second group, including endemic taxa, suggest the region to be the secondary center of their origin. Therefore, analyzing Tables 1–10, we may identify centers of this type. For instance, the Altai–Sayan region served as the secondary generation center of regular archaeocyathids during the early and middle Atdabanian, northern Africa in the middle Atdabanian, and Australia in the middle Botomian. Western North America was probably in this state during the early Botomian time. However, we cannot argue for this with any certainty, because the region is insufficiently studied.

All other regions, such as Mongolia, the Far East, western Europe, western North America, Central Asia, and China, served as areas of colonization for regular archaeocyathids over the Early Cambrian. The fauna replenishment took place here by way of immigrants from other regions.

Earlier, the secondary centers of origin were detected by using complex indices (Rozanov, 1980, 1984; Webbey, 1984). They coincide well with the centers that we identified by using our Tables 1–10. This demonstrates the workability, and even the advantage, of our approach as compared to others, as it is simple in procedure and based on easily accessible data.

A sharp increase in the abundance of genera constituting our third group is registered in all the regions at the early and middle Botomian times, and was indubitably related to the very wide transgression resulting from the Paleopangea breakup. These events facilitated

Table 10. Dynamics of fauna renewal in China

Time	Total number of genera	New genera				
		total number	first group	second group	endemic genera in two groups	third group
b_3	6	0	0	0	0	0
b_2	16	3	3	0	0	0
b_1	13	10	6	0	0	4
at_4	3	3	3	0	0	0

migration of different genera from one region to another. At the same time, they considerably extended ecological vacancies in all regions. Precisely this was responsible for the enormous increase in the abundance of Botomian genera, which were able to colonize quickly (in the geological sense) vast areas. Thus we may definitely conclude that the high abundance of early and middle Botomian genera constituting the third group was a response to the sea transgression.

This conclusion is not new, but important in its methodical aspect. Speculating on this matter, we wanted to show how it is possible to interpret quantitative relations between immigrated, cosmopolitan, and endemic genera.

FACTORS RESPONSIBLE FOR ENDEMISM OF REGULAR ARCHAEOCYATHIDS

In many studies on problems of paleobiogeography, the origin of endemic genera used to be considered as a sign of regional isolation. However, is this the only factor responsible for their origin?

Analyzing Tables 1–10, one may easily note two points. First, in all the regions, the greatest number of endemics came into existence at the Early and Middle Botomian substages. Second, some endemic genera appeared in the Siberian platform during the Tommotian Stage, and also in the Altai–Sayan region and northern Africa at the Middle Atdabanian Substage, that is to say, precisely at the moments, when these regions turned into secondary centers of origin.

Let us consider first the early Botomian peak of endemism in the regions. As is argued, the high degree of endemism at that time was a result of the isolation and provincialism of particular geoblocks (Debrenne *et al.*, 1989; Papeko *et al.*, 1992; Debrenne, 1992). Nonetheless, the high percentage of genera constituting the third group of the early Botomian taxa suggests that this isolation, assumed to have lasted over several chrons, was lacking. One should also bear in mind that we take for endemics only those taxa that never migrated outside their regions, and that in all the regions the high abundance of endemic genera is followed by a greater number of taxa of the third group. Thus, we should seek an additional or even alternative explanation for the appearance of endemic genera.

Here we should remember as well that the growing abundance of endemics coincided in time with the moments of highest diversity of the archaeocyathid fauna in all regions. This feature was typical of both the areas of colonization (the Far East, Mongolia, and western Europe), and of regions that were the secondary centers of origin. This apparently suggests that conditions of high competition stimulated the origin of very specialized genera, which were able to survive only in regional ecological niches, where they were more successful than other competing forms. We also noted earlier (Markov and Naimark, 1994; Naimark and Markov, 1994) that moments of intense diversification in different groups of marine invertebrates are regularly associated with a greater abundance of unstable, highly specialized genera displaying narrow geographic distribution. This chronic peculiarity of different faunas (sea urchins, bryozoans, athyridids, nautiloids, rugose corals, and archaeocyathids) seems to be one of the general features in the taxon's evolution. The appearance of endemic archaeocyathids at moments of their intense diversification well fits this regularity, if we consider endemism as an indicator of a high specialization of regional genera.

As noted above, endemics originated as well in the formation stages of secondary centers of origin. Even in the Siberian platform, which is the homeland for the most primitive and nonspecialized archaeocyathids, the Tommotian Stage is marked by the origin of two endemic genera. We tend to consider this event again as a result of a diminishing number of ecological niches and concurrently activated competition, as centers of origin are the very areas, where these niches are under permanent attack.

We also assume the active appearance of endemics in the Altai–Sayan region and northern Africa to be the result of the middle Botomian transformation of these areas into secondary centers of origin, while the next Botomian peak of endemism was related to a significant diversification of regional faunas. In the Australian center of origin, these two factors were active simultaneously. In contrast to the centers of origin, vacant ecological niches in areas of colonization were gradually occupied by taxa, and endemic genera appeared here later, when the fauna diversity was sufficiently high.

GENERALIZED SCHEME OF REGULAR ARCHAEOCYATHID EVOLUTION

As shown for nine regions, the history of regular archaeocyathid faunas begins with primitive nonspecialized genera *Archaeolynthus, Dokidocyathus, Ajacicyathus, Nochoroicyathus, Erismacoscinus*, and some others. All pioneer taxa in other regional faunas were born in the Siberian platform during the Tommotian Stage. They represented all suborders with the exception

of Tabulacyathina. Accordingly, regional diversification was within the taxonomic range from superfamilies to species, whereas the primary center of origin (Siberian platform) was the homeland for higher taxa, such as orders and suborders. The only exception to this rule is the origin of suborder Tabulacyathina in the Altai–Sayan area. However even here, the first representatives of regional fauna were immigrants from the Siberian platform, and at this evolutionary stage archaeocyathids included only primitive nonspecialized forms.

The second stage of archaeocyathid evolution is distinct due to the appearance of more and more specialized genera apparently replacing more "general" taxa in the course of competition. Their specialization was controlled by both the invasion of specialized immigrants, and the origin of specialized aboriginal genera. At this stage, studied regions experienced the incursion of archaeocyathids born in secondary centers of origin, such as the Altai–Sayan region, northern Africa, and Australia. These centers were favorable for endemic genera, and the later the center appeared, the more endemics originated in this area. For instance, the Australian fauna included 30% endemic taxa by the end of the Atdabanian time. In this particular case, we assume that adaptive space in other regions was already occupied at this time, and these areas had no vacant ecological niches for newly born genera, which therefore had to remain in their homeland. The growing degree of endemism, which is illustrated in circular diagrams displaying percentages of endemics relative to total number of genera (Figs. 1–6), is a result of progressive fauna specialization, but not a consequence of regional isolation.

This attracted our attention to the general trend depicting the progressive growth in both provincialism of faunas and number of biochores during the Paleozoic Era (Grunt *et al.*, 1994). In addition to suggested changes in paleogeography and geological situation over such a considerable time span, this trend could also be controlled by gradual specialization of the Paleozoic biota.

Different positions of the early, or middle Botomian diversity peaks are presumably related to the inequality of extinction rates in different regions. Considering above the chronology of extinction events in every region, we already noted that any displacement of maximum diversification toward a later period corresponds to a delayed commencement of extinction events. In its turn, the extinction rate is dependent on the average specialization of taxa in the region, which was not very high at the initial stage of fauna evolution. Due to this factor, the earlier start of fauna evolution may correspond to a later position of the maximum as with faunas of the Far East and northern Africa.

The third stage in the evolution of regional faunas of regular archaeocyathids is the period of their decline. This stage is marked by drops in general fauna diversity and by the extinction of all endemic genera represented by the most unstable and specialized regional forms. The few genera that participated in fauna renewal at this stage were newcomers from other regions, where they appeared in the middle–late Botomian interval.

Studying regional tetracoral faunas, Oliver and Pedder (1994) also noted some parallels in their evolution and synchronism of origin, or extinction events. These researchers arrived at the conclusion that some general factors independent of the regional position control, to a considerable degree, the evolution of tetracoral faunas.

Yet in some regions, the history of regular archaeocyathids deviates from the scheme outlined above. For instance, almost none of the new genera are recorded among residents of the Siberian platform at the middle Botomian time, when fauna diversity also decreased here. We believe that this was in response to the regional contraction of basin areas suitable for habitation of archaeocyathids.

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CONCLUSION

(1) Our procedure of subdividing archaeocyathid genera into three groups (aboriginal taxa, colonizers, and fast-migrating genera) is very convenient for detecting their centers of origin and areas of colonization.

(2) Endemic taxa may appear in the centers of origin and in all regions at moments of high generic abundance. There is a tendency for endemism to increase in regional faunas. In addition to all other factors, the endemism could be due to the increasing specialization of genera.

(3) The traced general scheme of regional taxa evolution includes three stages. The first of them corresponds to the introduction of primitive genera originating in the Siberian platform. The second stage consists of the gradual specialization of regional faunas at the expense of inborn taxa and other genera immigrated from secondary centers of origin. The third demonstrates the fauna decline, a significant drop in its diversity, the extinction of endemic and general taxa, and also a leveling of distinctions between regional faunas.

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On the Volume and Age of the Kapkatas Group of Northern Tien Shan

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Abstract—New data on the structure and age of the Kapkatas Group concern the lithostratigraphic volume of its formations and the first finds of diverse Lower Cambrian fossil assemblages. According to the hyolithid, hyolithelmintid, and protoconodont assemblages, the uppermost part of the group is correlative with the Botomian stratotype sections of the Siberian platform. The studied rock section differs from the Lower Cambrian sequence of tholeiitic basalts recorded in the Northern Tien Shan Caledonides in its basalt–andesite–dacite composition and calc-alkaline petrochemical trend.

Key words: *Northern Tien Shan, Kapka-Tash Range, Cambrian, Kapkatas Group, Techar Formation, fossil assemblages, small-shelled fossils, protoconodonts, brachiopods*

The Kapkatas Group was first defined by S.S. Shul'ts in 1936 as an independent stratigraphic unit of the Kapka-Tash Mountains under the name "Kapkatas Formation" of presumably Silurian age. The formation has been described to consist of diabase and andesite porphyrites, tuffaceous conglomerates and sandstones, and tuffites with limestone horizons.

Later, Krivolutskaya and Korolev (1960), who correlated "the Kapkatas Formation" with Cambrian spilite–keratophyre formations of Central Kazakhstan, proposed to treat it as a unit of the group rank. Simultaneously, these authors included into this group all the basalt–andesite and volcanogenic–sedimentary Caledonian sequences known then in Northern Tien Shan (the Song-Kel'-Too and Karakatty ranges, the northern slope of the Terskei Ala-Too Range, and the western part of the Kyrgyz Range).

The idea that deposits of this group are so widely spread, and the absence of paleontologic age constraints in the stratotype area, led later researchers to ambiguous interpretations of the group volume and, as a result, to the almost complete oblivion of this stratigraphic unit when elaborating recent stratigraphic schemes.

Seven units composing the Kapkatas Group in the stratotype section (Krivolutskaya and Korolev, 1960) were later grouped (*Dokembrii ...*, 1982; *Stratifitsirovannye ...*, 1982) into the Bel'tepshi, Sultan-Sary, and Techar formations (table and Fig. 1). Based on petrographical data, Dodonova (*Stratifitsirovannye ...*, 1982) showed that pillow basalts from the Kapka-Tash Range (the Bel'tepshi Formation) considerably differ in composition from basalts of the Karakatty Formation in the northern Son-Kul' area whose Early Cambrian age is supported by algae. These data and other geological evidence provided the basis for ascribing volcanics of the Kapkatas section to the Proterozoic. V.V. Kiselev and other researchers (*Dokembrii ...*, 1982), sharing this view on the Precambrian age of these deposits, suggested a paradoxical approach to solve this problem: to retain the Kapkatas Formation as a taxonomic unit of Cambrian age, and simultaneously to withdraw it from the stratotype section. Accordingly, the Bel'tepshi Formation was included in the Terskei Group whereas the overlying Sultan-Sary and Techar formations were proposed to be united into an independent group (*Dokembrii ...*, 1982). The stratigraphic scheme of the Kapka-Tash Range proposed by these researchers was based on their assumption that the Sultan-Sary Formation is completely composed of volcanics and that a hiatus exists at its base.

This view is inconsistent with our results confirming the validity of the Kapkatas Group (Krivolutskaya and Korolev, 1960) as a succession of lithostratigraphic units interrelated by similar formation conditions. We believe that paleontologic datings obtained for deposits of this group allow the question of its stratigraphic scope to be clarified. The complete section of the group is observed along left tributaries of the Sultan-Sary River, from the Buchuk Creek in the west to the Kumbel' Creek in the east (Figs. 1 and 2a). The lower part of the section is composed of pillow and tube basalts of the Bel'tepshi Formation up to 1000–1100 m thick. Remarkable among them are varieties with spilite, diabase, and hyalopilite textures. The outer rim of pillow lavas and volcanic tubes consists of amygdaloid rock varieties. Basalts are light-green, leucocratic, and show low titanium content. Lavas incorporate rare lenses of green and red cherts. These rocks exposed along the

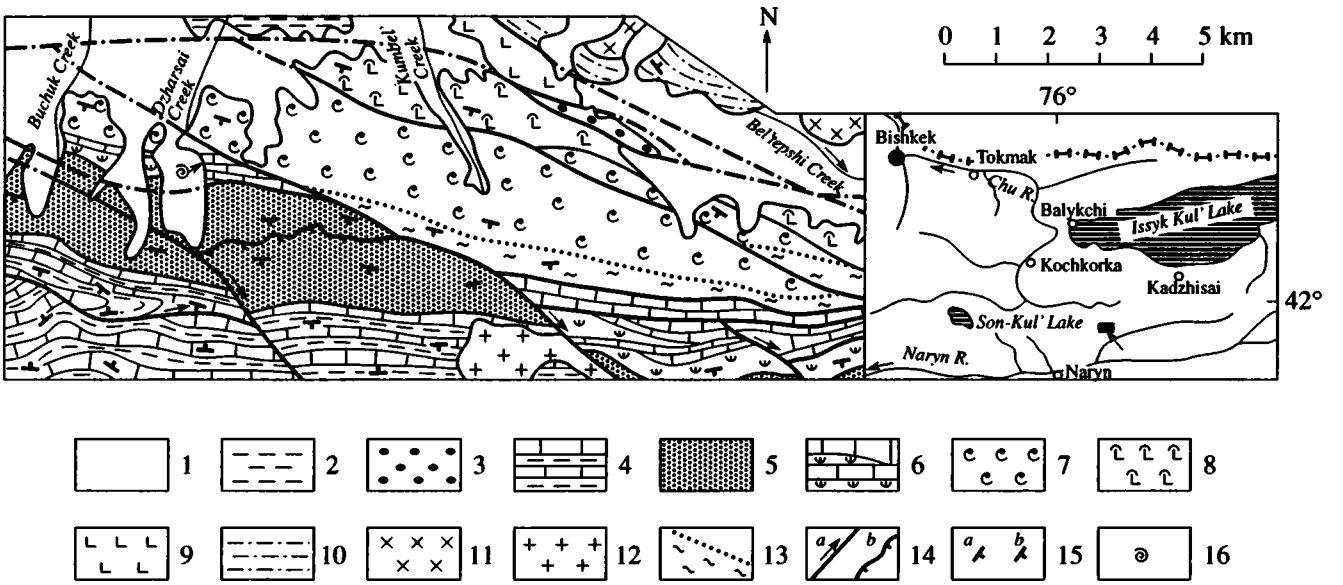

Fig. 1. Geological map of the western segment of the Kapka-Tash Range (compiled by A.V. Mikolaichuk): 1, Quaternary deposits; 2, Neogene deposits: marls, clays; 3, the Aktailyak Formation ($Є_2$): conglomerates, gritstones; 4, the Karagyr Formation ($Є_3$–O_1 ?): marmorized limestones, shales; 5, the Dzholdzhilga Formation ($Є_3$?): arkose sandstones, siltstones, shales; 6, the Techar Formation ($Є_{1-2}$): limestones, marbles, andesite-dacite tuffs; 7, the Sultan-Sary Formation ($Є_1$): basalt and andesite–dacite lavas and tuffs; 8, the Bel'tepshi Formation ($Є_1$): pillow basalt lavas; 9, the Karakatty Formation ($Є_{1-2}$): basalts, tuffs, cherts; 10, siltstones and dolomitic limestones (R_2?), 11, Ordovician granites and granodiorites; 12, Permian leucocratic granites; 13, dynamometamorphic schists, 14, faults: (*a*) steeply dipping, strike-slip faults included, (*b*) thrusts; 15, dip angles: (*a*) of sedimentary bedding and (*b*) of metamorphic lamination and cleavage; 16, occurrence of fossils.

Techar Creek bear sponge microscleres (Vasil'ev, 1991) known to exist since the Early Cambrian.

The overlying Sultan-Sary Formation mostly consists of pyroclastic material of a wide compositional spectrum. Most common are agglomerate, lapilli, and psammitic tuffs of andesites and dacites. Basaltic and basaltic andesite tuffs are present in subordinate amounts. The upper part of the formation is marked by the presence of tephritoids with graded bedding and rhyolite tuffs. Throughout the section, basalt, andesite, and dacite lava flows occur among pyroclastic rocks. Rocks reveal the tendency of silica content increase from the base of the section upward. The thickness of the formation is 2100 m.

The contact of these rocks with basalts of the Bel'tepshi Formation commonly coincides with a linear fault; the gradual replacement of Bel'tepshi basalts along the strike by pyroclastic deposits is observed only in the watershed area between the Techar and Kumbel' rivers. Thus, we place the lower boundary of the Sultan-Sary

Correlation of stratigraphic schemes of the northern slope of the Kapka-Tash Range (the Kapkatas Group)

Krivolutskaya and Korolev, 1960	Dodonova in (*Stratifitsirovannye...*, 1982)	Kiselev and others in (*Dokembrii...*, 1982)
Limestones, shales	Sultan-Sary Formation	Techar Formation
Tuffs, tuffaceous sandstones and shales		
Marbles		
Volcanogenic–sedimentary rocks: plagioclase and augite porphyrites, tuffs		
Chlorite–siliceous rocks: marmorized platty limestones, and chlorite–siliceous and tuffaceous schists		
Bedded tuffaceous conglomerates, psammitic tuffs, rare plagioporphyrite and diabase porphyrite flows		Sultan-Sary Formation
Volcanics: pillow basalts and andesitic porphyrite flows	Bel'tepshi Formation	Bel'tepshi Formation

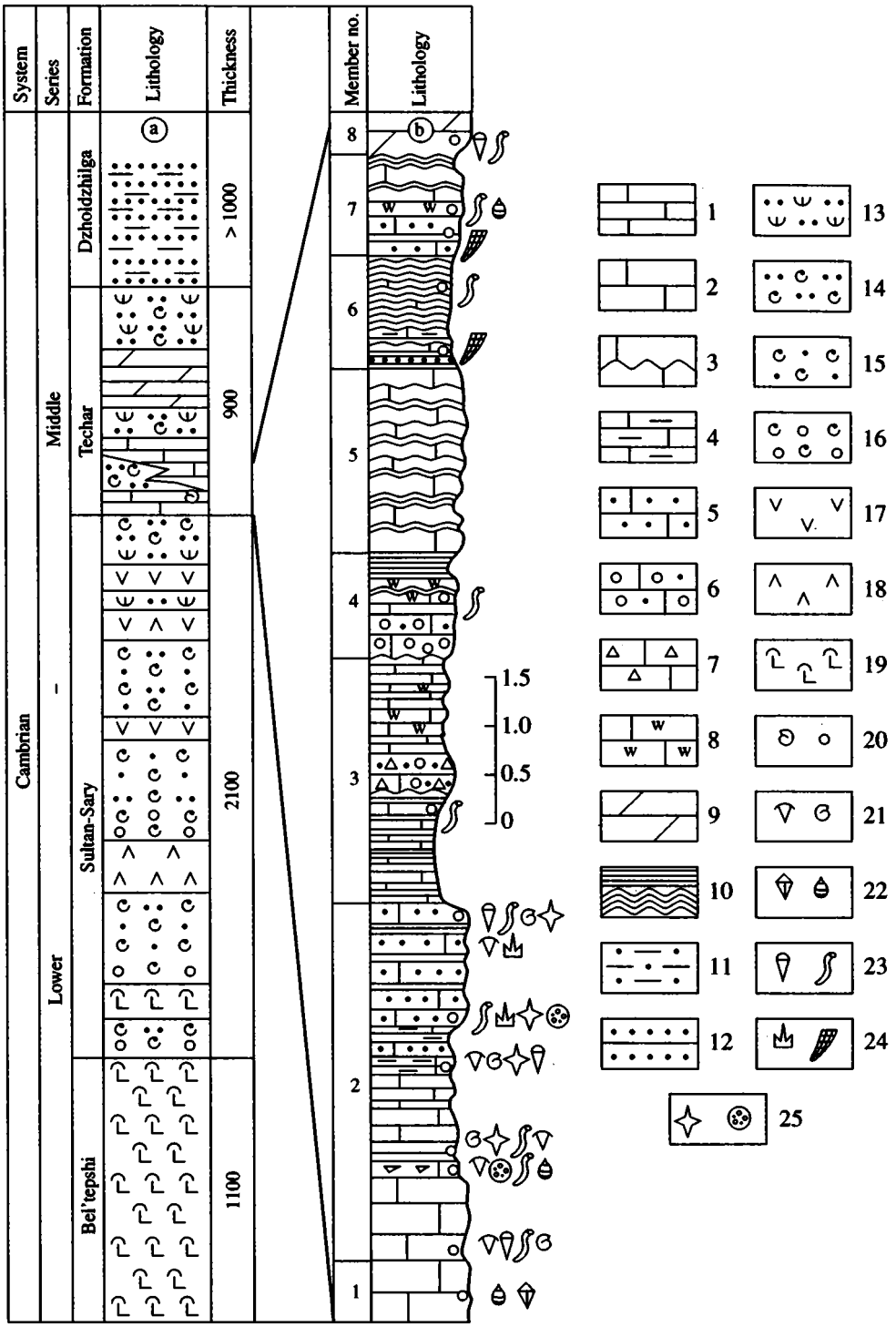

Fig. 2. Composite stratigraphic section of the Kapkatas Group (a) and beds of the Techar Formation (b): 1, thin- to medium-bedded limestones; 2, thick-bedded and massive limestones; 3, irregularly- and wavy-bedded limestones; 4, clayey limestones; 5, sandy limestones; 6, carbonate conglomerate breccia; 7, clastic brecciated limestones; 8, limestones with chert interlayers and lenses; 9, marbles and marmorized limestones; 10, shales; 11, siltstones; 12, sandstones; 13, tuffaceous sandstones; 14, psammitic tuffs; 15, lapilli tuffs; 16, agglomerate tuffs; 17, dacites; 18, andesites; 19, basaltic pillow lavas; 20, sampling places. Fossils: 21, trilobites and gastropods; 22, articulate and inarticulate brachiopods; 23, hyolithids and hyolithelmintids; 24, protoconodonts, conodonts, and tommotiids; 25, sponges and phosphate problematics.

Formation directly below the first occurrence of basaltic and andesitic tuffs in the section. Tuffs of the Sultan-Sary Formation are overlain by a 5- to 15-m-thick horizon of lithoclastic quartzose sandstones marking the base of the Techar Formation. Upsection, sandstones give way to marmorized limestones intercalated with gray clayey or organogenic-detrital limestones and green to lilac andesite–dacite tuffs. This well-preserved 290-m-thick lower part of the formation is exposed in the Buchuk–Dzharsai interfluve. Its upper layers are observed south of the Tarmaltor Pass where they are represented by pinkish and gray thin-banded marbles, psammitic tuffs of intermediate composition, and tuffaceous sandstones. Carbonate and volcanogenic–terrigenous rocks compose uniform members 50–200 m thick, which either alternate in the section, or laterally replace each other. The total thickness of the Techar Formation is estimated to make up 900 m. The formation is overlain with a sharp contact by arkose sandstones and siltstones of the Dzholdzhilga Formation.

Our predecessors repeatedly mentioned banded carbonate–chlorite and carbonate–chlorite schists and phillites in the Kapkatas Formation, the presence of which was treated as a feature favoring its Precambrian age. According to our observation, banded green schists, which associate with volcanics and their tuffs showing intense cleavage and boudinage structures, compose a series of latitudinal dynamometamorphic zones. In these zones, the banded structure is obliquely superimposed on the host rock bedding. The widest zone (up to 500 m) is traced along the watershed part of the Kapka-Tash Range near the contact between the Sultan-Sary and Techar formations over a distance of more than 15 km. To define the type of dynamometamorphism, special investigations need to be performed. Most likely, zones of dynamometamorphic schists and intense cleavage represent one of the elements of the structural paragenesis of the post-Caledonian folding.

A section of the Techar Formation of the Kapkatas Group bearing abundant micro- and macrofossils was studied in detail along the Dzharsai Creek. In the right bank of the creek, psammitic dacite tuffs of the Sultan-Sary Formation are overlain by lithoclastic quartzose sandstones 15 m thick. Upsection, a 7.5-m-thick bed of brownish-yellow clayey limestones underlies the following succession:

(1) Marmorized limestones 30–35 m thick (Fig. 2b, Member 1). They are represented by gray and dark-gray varieties with thick and obscure bedding. Ten meters above the base of the member, there is a thin layer (up to 5 cm) bearing single articulate brachiopods *Nisusia* sp. The upper layers yielded single hyolithelmintid (worm tubes) remains determined as *Hyolithellus* sp. Rare fragments of inarticulate brachiopod shells also occur.

(2) Gray and dark-gray medium- to thick-bedded sandy limestones (Fig. 2b, Member 2) with crimson-colored shales and clayey limestone interlayers. Locally, limestones become thin-banded; clastic-brecciated varieties occur at the base of the member. Rare thin layers of aphanitic limestones contain trilobite and brachiopod remains. Various fossils collected bed-by-bed at five stratigraphic levels are trilobites of the *Olenoides* genus, articulate brachiopods of the *Nisusia* genus and inarticulate brachiopods of the *Linnarssonia*, *Prototreta*, and *Homotreta* genera. Abundant hyolithid, hyolithelmintid, gastropods, ostracodes, protoconodonts, and sponges occur throughout the member. The most representative fossil assemblage was recorded in the top layers. It includes: (1) hyolithid *Microcornus parvulus* Mamb., *M. talasicus* Mamb., *Contitheca teslenkoi* Mamb. et Meshk., *C.* sp., *Adyshevitheca* cf. *adyshevi* Mamb., *Burithes* ex gr. *elongatus* Miss., *Laticornus* sp., and *Sulcavitidae* gen. et sp. indet.; (2) hyolithelmintid *Hyolithellus vladimirovae* Miss., *H. mikolaitchukii* sp. nov., *H.* cf. *micans* Billin., *H.* sp., *Torellella grandis* Meshk., *T.* cf. *biconvexa* Miss., and *T.* sp.; (3) gastropods *Latouchella accordionata* Runn. et Jell., *Jochelcionella* cf. *recta* Miss., *Mellopegma* cf. *georgiensis* Runn. et Jell., *Shabaktiella* cf. *shabaktiensis* Miss., *Pelagiella lorentzi* Kob., *P.* sp., and *Jgorella* cf. *atdabanica* (Miss.); (4) protoconodonts *Protohertzina sicciformis* Miss., *Amphigeizina renae* Mamb. et Miss., *Hertzina*(?) *bisulcata* Mull., and *Jsoxus* sp.; (5) celoscleritophorids *Chancelloria* cf. *maroccana* Sdzny; (6) sponge spicules *Eiffelia araniformis* (Miss.); (7) problematics *Microdictyon effusum* Bengt., Matth. et Miss.; (8) tommotiids *Tesella* sp.; (9) ostracodes *Bradoria* sp.; and (10) some other problematics not yet described in the literature. The thickness of the member is 3.8–4.0 m.

(3) Dark-gray thin-bedded limestones interlayering with dark crimson shales (Fig. 2b, Member 3). One meter above the base, a bed of carbonate gritstones (up to 0.3 m) occurs. Rare layers and lenses (up to 10 cm) of black cherts are present among limestones. The thickness of the member is 2.5 m.

(4) Rocks of Member 3 are overlain above an uneven erosional surface by dark-gray gritstones and sandstones, which are replaced upsection by thin-bedded pinkish-gray and crimson shales with black chert interlayers (Fig. 2b, Member 4). Chemical maceration of rock samples from some intervals yielded remains of inarticulate brachiopods and hyolithelmintid *Hyolithellus* sp. The thickness is 0.8–1.0 m.

(5) Recrystallized and intensely fractured medium-bedded limestones with thin interlayers of dark crimson shales (Fig. 2b, Member 5). The thickness is 2 m.

(6) Dark crimson fractured shales (Fig. 2b, Member 6) with thin (up to 5 cm) interlayers and lenses of yellowish-brown clayey limestones. The base of the member is marked by sandy limestone beds with protoconodonts *Rhombocorniculum* ex gr. *undulatum* Miss. and tommotiids *Tesella* ex gr. *trilateralis* Miss. et Grig. Single remains of *Huolithellus* sp. occur throughout the member. The thickness is 1.2 m.

(7) Gray and pinkish-gray sandy limestones 40 cm thick occur in the lower part of the member (Fig. 2b, Member 7). Abundant remains of tommotiids extracted by maceration from these rocks are *Tesella* cf. *deplanata* Miss. et Grig. and *T.* ex gr. *trilateralis* Miss. et Grig. Upsection, limestones became bluish-gray and incorporate dark crimson shale interlayers and black chert bed at the base. Rare inarticulate brachiopods were recorded. The thickness is 0.6 m.

(8) Dark-gray marmorized limestones with thick, occasionally obscure bedding and a dense net of calcite veinlets (Fig. 2b, Member 8). Thin-bedded limestones appear 5.5–6.0 m above the base. In the top part, brownish-gray, intensely foliated tuff layers (up to 0.8–1.0 m) occur. The lower layers bear single *Sulcavitidae* gen. et sp. indet., *Hyolithellus* ex gr. *micans* Bill, and indeterminable fragments of inarticulate brachiopods. The thickness is 11.8–12.0 m.

The upper part of the sequence is composed of light-gray and pinkish massive and thin-bedded marmorized limestones with 25- to 35-m-thick horizons of tuffs of intermediate composition. This portion of the section lacks organic remains. In the south, limestones border sandstones and silty shales of the Dzholdzhilga Formation along the sublatitudinal fault.

Fossils from the Dzhartash section were determined by G.Kh. Ergaliev (trilobites), P.P. Misyus (articulate and inarticulate brachiopods), and A.M. Mambetov (ostracodes, hyolithid, hyolithelmintid, conodonts, protoconodonts, and phosphate problematics).

Articulate and inarticulate brachiopods found in the Techar Formation are represented by new species of the *Nisusia* and *Linnarssonia* genera. Representatives of the *Nisusia* genus are widely spread in the Lower–Middle Cambrian deposits of North America and Argentina (*Treatise* ..., 1965), and central Kazakhstan, where they were recorded in layers referred to the Maidan Horizon (Romanenko *et al.*, 1967; Koneva, 1979), and also in the lower Middle Cambrian (the *Oryctocara* Zone) of the Altai–Sayan region (Aksarina, 1960; Romanenko *et al.*, 1967; Aksarina and Pel'man, 1978). Species of the *Linnarssonia* genus display the same stratigraphic range (Walcott, 1902; Schuchert and Cooper, 1932; *Treatise* ..., 1965; Romanenko *et al.*, 1967; Koneva, 1979). In addition to brachiopods, limestones of the Techar Formation enclose abundant trilobites of the *Olenoides* genus reported from Lower–Middle Cambrian deposits of North America and Central Asia (*Treatise* ..., 1959, 1965).

Brachiopods and trilobites found in the Dzartash section are represented by new endemic species; therefore, when defining the age of host rocks, we used the assemblages of small-shelled fossils.

Small-shelled fossils chemically extracted from rocks of the same section belong to two biostratigraphic assemblages. The first of them is characteristic of Members 1 and 2 of the above-described section (Fig. 2b) and represented by such fossils as: (1) hyolithid *Microcornus parvulus* Mamb., *M. talasicus* Mamb., and *Adyshevitheca* cf. *adyshevi* Mamb.; (2) hyolithelmintid *Hyolithellus vladimirovae* Miss. and *Torellella grandis* Meshk.; (3) gastropods and rostroconch mollusks *Latouchella accordionata* Runn. et Jell., *Yochelcionella* cf. *recta* Miss., *Mellopegma* cf. *georgiensis* Runn. et Jell., *Pelagiella lorenzi* Kob., and *Igorella* cf. *atdabanica* (Miss.); (4) protoconodonts *Protosertzina sicciformis* Miss., *Amphigeisina renae* Mamb. et Miss., and *Hertzina*(?) *bisulcata* Mull., and (5) sponges *Eiffelia araniformis* (Miss.). Similar forms were recorded in the Siberian platform, Central Tien Shan, and southern Kazakhstan (Mambetov and Repina, 1979; Mambetov and Imanaliev, 1981; Missarzhevskii and Mambetov, 1981; Missarzhevskii, 1989), and also in China, Iran, and Australia (Mambetov and Repina; 1979; Yu Wen, 1979; Missarzhevskii, 1989; Hamdi, 1989; Qian Yi and Bengtson, 1989; Bengtson *et al.*, 1990) from deposits correlated with the Botomian Stage.

The second assemblage found in Members 4–8 of the Dzharsai section is much less diverse at both the species and generic levels. It consists of representatives of the *Rhomlocorniculum* (protoconodonts) and *Tesella* (tommotiids) genera. Single hyolithid (the Sulcavitidae family) and hyolithelmintid (the *Hyolithellus* genus) are also present. The analysis of the vertical distribution of fossils reveals that host deposits correspond to the uppermost Botomian Stage in stratotype sections of the Siberian platform. For instance, *Tesella* forms appear in the uppermost Botomian (Lower Cambrian) to lowermost Amgan (Middle Cambrian) interval whereas the species *Rhomlocorniculum ungulatum* Miss first occurs in the lowermost Botomian Stage and does not cross the Lower–Middle Cambrian boundary (Missarzhevskii, 1989).

Whereas the second assemblage from the Cambrian sequences of Tien Shan was formerly unknown and requires further study, the first fossil assemblage from the Dzharsai Creek section is identical with that which was earlier distinguished in the Malyi Karatau Range as characteristic of the *Microcornus parvulus* Zone (Mambetov and Repina, 1979) correlative with the lowermost Botomian Stage of the Siberian platform (Missarzhevskii, 1989). Later, this zone was traced through the Talas–Karatau (the Talas Ala-Too Range) and further, in the Chatkal–Naryn (the Chatkal Range) structural zones of Tien Shan (Mambetov and Repina, 1979; Mambetov and Imanaliev, 1981; Missarzhevskii and Mambetov, 1981). Thus, recent data suggest that Cambrian sections of northern and central Tien Shan (i.e., located, according to available tectonic schemes, in different structural zones) represent, in fact, elements of a formerly single biogeographic province.

Paleontologic dating of the Kapkatas Group also turned out to be important for solving more general problems of regional geology. According to an oral communication of I.L. Zakharov and M.P. Khristova as well as other data (Vasil'ev, 1991), two rock associa-

tions differing in their petrochemical trends can be distinguished among Cambrian–Ordovician volcanics of northern Tien Shan. One of them consists of tholeiitic basalts and is a part of the ophiolite rock assemblage. Its age is reliably substantiated by data from the Karakatty Formation (Maksumova *et al.*, 1988; Mambetov and Apayarov, 1990). The second rock association (it includes also the Bel'tepshi Formation) features differentiated basalt–andesite–dacite series of calc-alkaline affinity. Until recently, its fossils were represented only by single sponge spicules, which were spatially separated and, therefore, their association with differentiated rock was called into question. We are inclined to consider the presented materials as a missing argument in favor of the lateral petrochemical heterogeneity of Lower Paleozoic volcanogenic sequences in the Caledonides of northern Tien Shan.

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Radiometric Age of Hyperacidic Volcanism in Late-Orogenic Troughs of the Chatkal–Kurama Zone, Central Tien Shan

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Abstract—The number and age of hyperacidic volcanism events in late-orogenic troughs of the Chatkal–Kurama structural zone are discussed on the basis of the radiometric dating. All K–Ar data available are used, including 90 original age determinations. The results of Rb–Sr dating obtained for five liparitic and trachyliparitic volcanic bodies of the region are represented. The reliability of the radiometric data is analyzed. The origin of most hyperacidic volcanic bodies of the region is shown to occur in Early Permian (Asselian). The occurrence of younger (P_2–T_1) acidic volcanics, which is suggested in the traditional scheme of the evolution of volcanism in the region, is not confirmed by radiometric data. The materials presented suggest pre-Asselian (C_2?) hyperacidic volcanics in the region.

Key words: *radiometric dating, volcanics, orogenic troughs, Central Tien Shan*

FORMULATION OF THE PROBLEM

Regularities in the evolution of orogenic belts still attract the attention of geologists. In particular, one of the most interesting problems is the evolution of orogenic volcanism, which can be revealed only through investigation of the genetic succession of the corresponding volcanic bodies. In many cases the succession can be established only on the basis of age determinations. However, the possibility of paleontological dating in the sequences, which accumulated during orogenic events, is very limited because they bear scarce fossils. Therefore, in studies on the stratigraphy of orogenic areas, the radiometric dating of rocks is of prime importance. This work represents an attempt to establish the number and age of hyperacidic volcanic pulses in late-orogenic troughs of the Chatkal–Kurama structural zone, which underwent an orogenic stage in the Late Paleozoic. So far these problems have found no adequate solution, though geological studies in the region have lasted nearly half of our century.

Conclusions on the position of hyperacidic volcanics among other volcanic series of this region depend largely upon the disputable age of the trachyliparite Kyzylnura Formation, which is widespread here. Disagreements on its age arose at the beginning of the 60s, when Laverov *et al.* (1964) presented ten K–Ar age determinations corresponding to the Early Permian for the Babaitaudor subvolcanic massif, which is unanimously recognized as the type of the Kyzylnura Formation. These data appeared to be in conflict with the traditional opinion formed in the course of correlation of volcanic sequences (usually paleontologically barren) from different, spatially separated, and often remote late-orogenic troughs of the Chatkal–Kurama zone. According to this view, the Kyzulnura Formation is assumed to be Early Triassic in age (Vasil'kovskii, 1952); this was later confirmed by remnants of terrestrial flora from deposits of this formation (the Bashkyzylsai River valley), which was assigned to the Early Triassic or Late Permian (Sikstel', 1960).

It appeared impossible to reconcile radiometric and floral datings for the Kyzylnura volcanics, because the K–Ar age determinations for liparites, overlying the horizon with fossil flora, assigned to the Triassic, suggest them to be as old as the Early Permian (264 ± 10 Ma)[1] (Belikov *et al.*, 1964). Thus geologists had to choose the method to be used in their concepts.

Publications on geology of the region show that so far many geologists actually ignore the K–Ar data on the Kyzylnura Formation and hold the traditional view of its age (Arapov, 1983; Babakhodzhaev *et al.*, 1993; *Geologiya SSSR*, 1972; Gleizer *et al.*, 1976; *Petrografiya Tadzhikistana*, 1986; *Raschlenenie ...*, 1976; Tkachev *et al.*, 1984; Tulyaganov *et al.*, 1984; Shayakubov *et al.*, 1988). The stratigraphic schemes

[1] K–Ar and Rb–Sr ages presented in this work are estimated using the decay constants recommended by the International Subcommission on Geochronology.

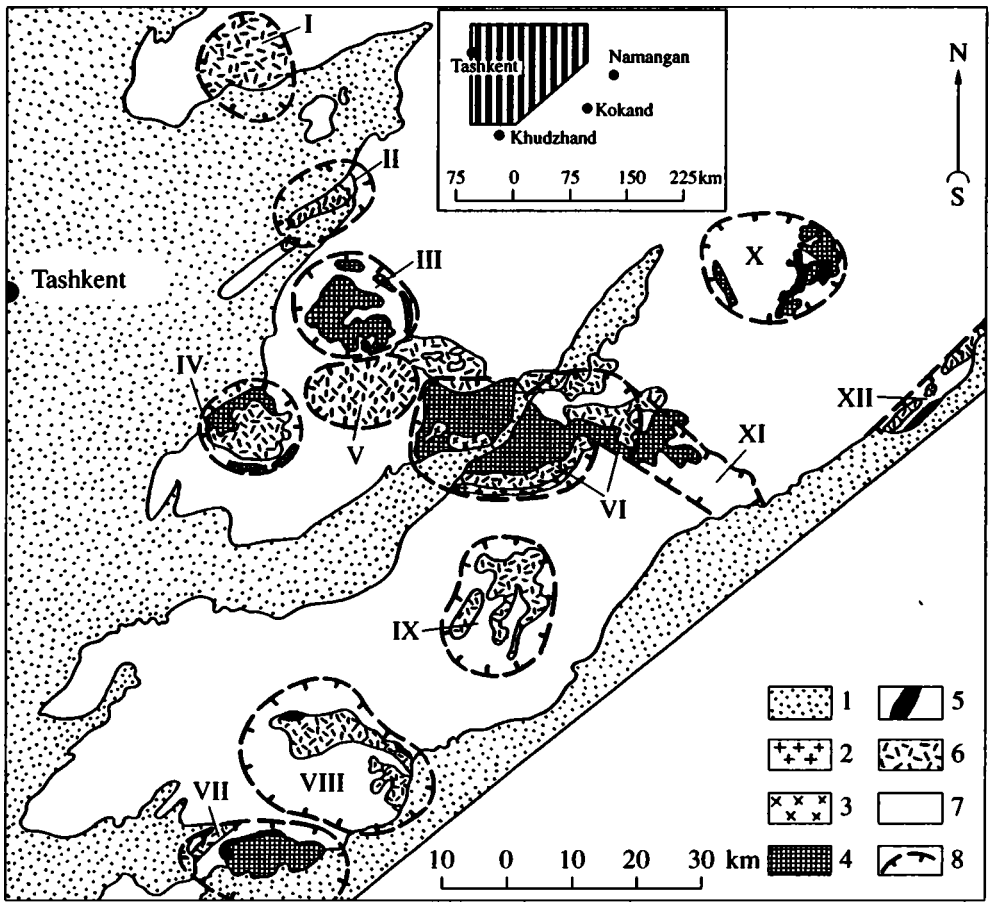

Fig. 1. Exposures of late-orogenic hyperacidic volcanics in volcano-tectonic structures of the Chatkal–Kurama zone. Classification and position of structures after Arapov (1983). Calderas: I, Oyasai; II, Maigashkan; III, Kyzylnura; IV, Karabash; V, Chilten; VI, Babaitaudor. Basins: VII, Tashkesken; VIII, Adrasman; IX, Lashkerek; X, Kugaly. Fault troughs: XI, Chadaksai; XII, Kumlai-Koksarek. 1, Mezo–Cenozoic deposits; 2, Muzbel'-type granite-porphyry; 3, Babaitaudor-type granosyenite-porphyry; 4–6, hyperacidic volcanics: 4, Kyzylnura Formation, 5, Ravash Formation, 6, Oyasai Formation; 7, other Paleozoic complexes; 8, boundaries of late-orogenic troughs.

reflecting this viewpoint set the Kyzylnura volcanics against other hyperacidic volcanics united into the liparite Oyasai Formation (Fig. 1), which occur in the same troughs. Early Permian terrestrial flora occurs in many sequences attributed to the Oyasai Formation. The fossils of Asselian fusulinids were collected from the basal horizon of one of its sequences (Arapov *et al.*, 1978; Bensh *et al.*, 1982). This has led geologists to assign the Oyasai Formation to the Asselian Stage of the Early Permian. It is assumed that between the Asselian and Early Triassic stages of acidic volcanism there are trachyandesite Shurabsai and trachybasalt-trachyliparite Ravash formations, which originated in the Sakmarian and Artinskian or Kungurian, respectively. In summary, according to the traditional view, the hyperacidic volcanism in Late Hercynian troughs occurred three times and produced volcanic bodies of the Oyasai, Kyzylnura, and partly of the Ravash Formations (Table 1).

The suggestion of the Early Permian age of the Kyzylnura Formation received support from Tadzhibaev (1979), and in addition, Dalimov and Rafikov (1986) presented several new arguments in its favor.

Dalimov and Rafikov believe that volcanics of the Oyasai and Kyzylnura Formations were formed at the same time and should be regarded as a single formation. Table 1 compares the regional scheme of volcanism suggested by these geologists with that based on the traditional concept.

To support the age compatibility of the Oyasai and Kyzylnura Formations, Dalimov (1981) noted the absence of any sections with both acidic volcanic sequences in the region. According to the traditional stratigraphic scheme, these two units should be separated by the trachyandesitic and trachybasaltic sequences. Dalimov also referred to the identity of the K–Ar dating ranges obtained for both formations (250–300 Ma). From Dalimov's viewpoint, their synchronism is also evident from the fact that deposits of the Kyzylnura Formation overlie the Oyasai volcanics

without stratigraphic evidences of a long gap between these volcanic deposits.

Although the first and last aforementioned arguments shed doubt on the idea of different ages for the Oyasai and Kyzylnura volcanics, they do not exclude this possibility. As for the K–Ar data, they can neither prove nor rule out this viewpoint because of the very wide range of data unverified by the reliability analysis.

Nevertheless, the publications reviewed above show that the traditional stratigraphic scheme of Late Paleozoic volcanic deposits has significant disadvantages, as it is based on methodologicaly imperfect investigations.

In our opinion, the main weak point of the above studies is that deposits filling up spatially separated troughs of the region were correlated on the basis of the lithologic and petrographic similarity of the sequences, and on their comparable stratigraphic position relative to paleontologically dated horizons.

Evidently this approach rests on the assumption that volcanism of the same type occurred simultaneously in all troughs of the Chatkal–Kurama zone. Moreover, this methodology requires at least one paleontologically dated horizon to be present in each correlated section. However, the last requirement is invalid in most cases, and until now the synchronism of comparable volcanic strata in the separated troughs has not been proved by real data, and has even been doubted (Borisov *et al.*, 1988; Vinichenko and Kukhtikov, 1985). It is also remarkable that volcanics of similar appearance and composition occur here in different parts of the stratigraphic section, and the lithological correlation usually results in two or even more alternative versions of stratigraphic relationship. Hence it is typical for the region that different geologists show in their maps the same rock unit as attributed to different formations with incompatible age indices.

One more shortcoming of the stratigraphic models in question is the low reliability of the paleontological data used. Indeed, the index marine fossils are found in only three out of two or three hundred Late Paleozoic units distinguished in the region. In the other cases the rock ages are evaluated on the basis of terrestrial paleoflora. As is generally known, paleobotanical data are less reliable, and in our case the reliability of floral dating seems to be especially doubtful as the degree of investigation of the Late Paleozoic flora in the region is low. The collected fossil plants are often poorly preserved and can be identified only at the generic level. The list of genera and species found is usually very short; many of them are endemic, and their time ranges are wide.

Active polemics about the age (C_2 or P_1?) of paleoflora collected from the Kandyrsai, Karabau, Baritovaya Gorka and other localities, shows that many published paleoflora datings for sediments filling in the orogenic troughs of the region are controversial (Arapov, 1983; Borisov *et al.*, 1988; Masumov *et al.*, 1978; Masumov, 1987; Tulyaganov *et al.*, 1984).

Table 1. Succession in formation of Upper Paleozoic volcanic formations of the Chatkal–Kurama Zone and their ages

After N. P. Vasil'kovskii (traditional scheme)		After T. N. Dalimov and Ya. M. Rafikov	
age	formation[1]	age	formation[1]
T_1	Kyzylnura		
P_1	Ravash	P_1–P_2	Ravash
P_1	Shurabsai	P_1	Shurabsai
P_1	Oyasai	P_1	Oyasai and Kyzylnura
C_{2-3}	Nadaksai	C_2	Akchasai and Nadaksai
C_2	Akchasai		
C_2	Minbulak	C_2	Minbulak

[1] Currently some geologists who work in this region prefer to designate a combination of synchronous stratified volcanic (volcanosedimentary), and comagmatic subvolcanic bodies as volcanic complexes rather than formations, preserving previous geographic names.

The accuracy of paleobotanical dating of the Kyzylnura Formation in the Bashkyzylsai valley should also be doubted, because fossil flora (*Equisetites* sp., *Williamsonia artemovae, Pseudovoltzia liebeana*, and *Angrenia angustifolia*) from this formation is, in the opinion of some paleontologists (Dobruskina, 1980), inappropriate for age determination.

Thus it is evident that available geological data and previous age constraints are inadequate to reliably solve the indicated problem. Therefore, additional reliable information on the age of the acid volcanics assigned to the different formations of the region is required. The possibility to get this information through the search and study of fossils appears to be limited, as has been shown by regional geological experience. This potential, in our view, has been exhausted. Thus the solution to the problem must be based on radiometric age determinations for appropriate volcanic rocks.

Accordingly in our geochronological investigations, we summarized data of other researchers and performed new isotopic age determinations for volcanics from the late orogenic troughs of the Chatkal–Kurama zone. We also regarded the degree of reliability of the total data set on isotropic ages as the most important aspect of our investigations.

RADIOMETRIC-GEOCHRONOLOGIC TECHNIQUE

To carry out the isotropic-geochronologic investigations, we used K–Ar and Rb–Sr techniques.

We sampled volcanic rocks from blocks that do not show any sign of weathering and local hydrothermal alteration. Samples for the isochron Rb–Sr analysis varied in mass from 8 to 10 kg. In the remaining cases the mass was not less than 1 kg. We measured Ar-, Rb-, and Sr-contents by the isotope dilution technique. The K-content was determined by flame photometry. Isoto-

pic analyses were performed on MI-1201 (Ar) and on MI-1320 (Rb and Sr) mass spectrometers. We controlled the accuracy of measurements by the international standard samples of muscovite P207 and biotite 70 of Eimer & Amend. The analytical error of K–Ar data does not exceed 2%. The error of the Sr and Rb radiometric analyses is 0.01 and 0.2% respectively. All the errors shown in Tables 2 and 3 correspond to a confidence level of 95%. The parameters of Rb–Sr isochrons are calculated by York's method. A detailed description of the analytical technique used was published previously (*Kriterii nadezhnosti ...*, 1976; *Mass-spektrometriya ...*, 1983).

RESULTS OF ISOTOPIC DATING

K–Ar age determinations. By now these data are available for acidic volcanics from most of the late orogenic troughs of the region. The compiled sample includs 120 data (Fig. 2). We considered only data obtained by the isotope dilution technique for $^{40}Ar_{rad}$ and by flame photometry for K. Sixty two data are selected from published works (Arapov *et al.*, 1978; Gleizer *et al.*, 1976; *Katalog opredelenii ...*, 1972; Laverov *et al.*, 1964). The other data are our own and published for the first time (Table 2). In addition, we use 69 K–Ar data for rocks and minerals from other geological formations of the region (Fig. 3). Thirty two of them are original and are also presented in Table 2. We performed analyses using both the whole-rock volcanic samples and the monomineral fractions of sanidine, orthoclase, and biotite from phenocryst assemblages and matrix of these rocks.

Histograms (Fig. 2) show the distribution of K–Ar age values for the acidic volcanics of the region. According to the traditional stratigraphic scheme, these volcanics belong to the Oyasai, Ravash, and Kyzylnura formations. Figure 3 displays a comparison of these and K–Ar data for rocks from other geological formations of the region, whose age relations with acidic volcanics are known from field evidences. In plotting the histograms, we mainly used the formation affinity of the dated volcanic bodies according to Arapov (1983). We accepted the viewpoint of Tulyaganov *et al.* (1984) only for the acidic volcanics of the Kumlai–Koksarek trough, and similar volcanics of the Tashkesken and Adrasman basins were classified according to the opinion published in *Petrografiya Tadzhikistana* (1986).

The histograms (Fig. 2) show the following distribution of age values.

For all three formations the K–Ar data vary over very wide and almost identical ranges: from 240 to 320 Ma for Oyasai and Ravash formations, and from 240 to 300 Ma for the Kyzylnura Formation.

The data distribution for the Kyzylnura formation (64 datings in total) is close to normal, with the most frequent values in the range from 260 to 270 Ma. This type of distribution is traceable in all Late Paleozoic troughs of the region. We should emphasize that the largest values for this formation are determined on sanidine phenocrysts and on high-K biotite without signs of metamorphic alteration (Fig. 3).

The data for the Ravash Formation (14 values in total) fall into two separate intervals of: 280–310 and 250–270 Ma. The first one was obtained on biotite phenocrysts without metamorphic alteration and the second on the whole-rock volcanic samples (Fig. 3).

The age histogram for the Oyasai Formation (42 analyses in total) outlines two maxima within the intervals of 250–280 and 310–320 Ma. This bimodal distribution cannot be related to the difference in mineral composition of the dated samples, because all data are obtained on the whole-rock samples of nearly the same composition. Moreover, the K–Ar age values in this collection depend on the sampling site: the older values are obtained only for volcanics of Karabash and Chilten calderas and are absent among data for the other sites of the region.

Rb–Sr age determinations. We pioneered the Rb–Sr method for dating the late orogenic acidic volcanics of the region. In total, we obtained five Rb–Sr isochron age values (Table 3, Fig. 4). Two of them characterize the age of volcanic bodies which are believed to be typical representatives of the Kyzylnura Formation (Babaitaudor subvolcanic massif and Samgar extrusion). One value is for post-Kyzylnura volcanic rocks (Muzbel'-type granite-porphyry from the Mailikotan dike). Two more ages are measured for volcanics of the Lashkerek and Tashkesken Members, which represent respectively the Oyasai and Ravash (or Kyzylnura according to some authors) formations. To construct isochrons, we used the analyses of both whole-rock volcanic samples and their constituents such as phenocrysts and coexisting matrices.

Studying the Babaitaudor massif, we carried out Rb–Sr analyses only for rocks of the first stage of its formation distinguished by geologists, who studied the massif in most details (Kantsel' *et al.*, 1972). We did this because not all investigators accept the rocks of the second stage (Babaitaudor-type granosienite-porphyries, Muzbel'-type granite-porphyries, and others) as belonging to the Kyzylnura Formation. We determined the age of liparites and granite-porphyries representing different phases of the first formation stage of the massif. According to our observation, the number of the phases is greater than was thought by previous researchers. We analyzed twelve samples in total. Four of them represent whole-rock volcanic material, two are a fine-grained matrix, and six are feldspar phenocrysts. The last fractions include transparent sanidines and plagioclase, and also orthoclase replaced by pelitic aggregate and dyed by iron hydroxides (Table 3, Fig. 4a). The age of the rocks calculated from the isochron parameters is 284 ± 1 Ma, the initial $^{87}Sr/^{86}Sr$ ratio is 0.70627 ± 15.

The isochron for the Samgar extrusion is based on four whole-rock samples of volcanics and two samples of fresh biotite phenocrysts from these rocks (Table 3, Fig. 4b). According to Tadzhibaev (1979), all rocks of this extrusion (alaskites, alaskite-porphyries, trachyliparites, and pegmatites) are comagmatic and formed at one time, although representing divergent facies. The isochron age of the extrusion is 281 ± 8 Ma, the initial $^{87}Sr/^{86}Sr$ ratio is 0.7051 ± 12.

The isochron for the Lashkerek Member (Fig. 4c) is plotted on the basis of eighteen samples, seven of which are whole-rock material of ignimbrites, one represents phenocrysts of pink orthoclase, and others consist of fiamme material and mineral segregations that compose bands and lenses in the ignimbrites. The analytical results were published earlier (Volkov *et al.*, 1990). The isochron age of the member is 284 ± 3 Ma, the initial $^{87}Sr/^{86}Sr$ ratio is 0.7066 ± 4.

The Tashkesken Member was dated by using mineral components of vitrophyric trachyliparites collected from the horizon of the so-called obsidians. This horizon is localized in the upper part of the member section. We analyzed samples of transparent oligoclase-andesine and sanidine that occur in the rocks as phenocrysts, and a sample of strongly hydrated volcanic glass of the rock matrix, five samples in total (Table 3, Fig. 4d). The isochron age of the rocks is 275 ± 4 Ma, and the initial $^{87}Sr/^{86}Sr$ ratio is 0.7070 ± 2.

The Rb–Sr isochron age of the Muzbel'-type granite-porphyries is determined to be 268 ± 5 Ma. These granite-porphyries are dominant and the latest rocks of the Mailikotan dike. Among the other rocks of the dike there are analogues of liparites of the Babaitaudor massif. We obtained the isochron through the study of one whole-rock granite-porphyry sample, two samples of brick-red orthoclase composing large phenocrysts replaced by pelitic aggregate, and two samples of micrograined matrix (Table 3, Fig. 4e). The initial $^{87}Sr/^{86}Sr$ ratio of the examined rocks and minerals is 0.7111 ± 15.

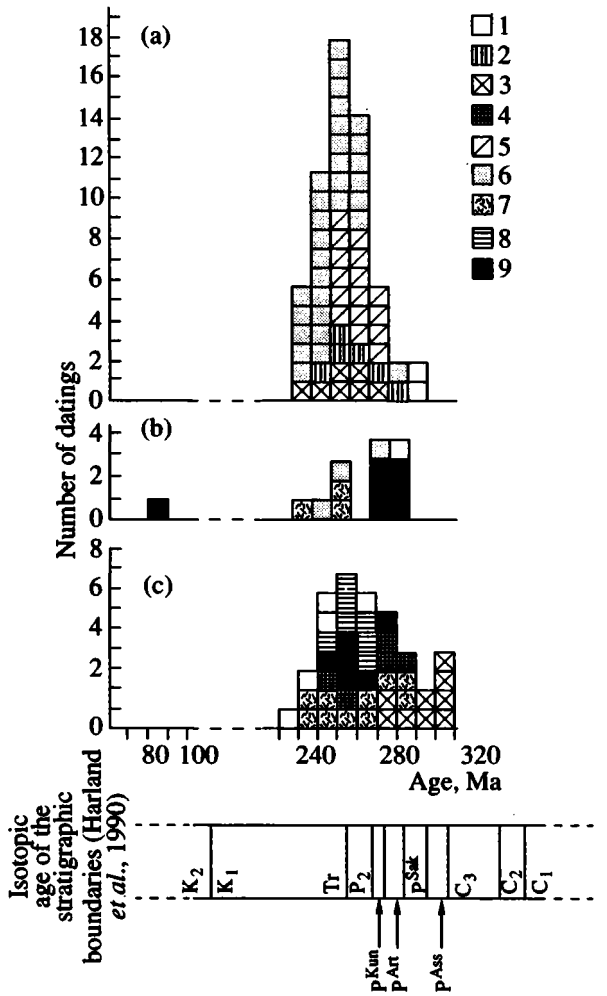

Fig. 2. Histogram of K–Ar ages for acid volcanics of the Kyzylnura (a), Ravash (b), and Oyasai (c) formations: 1–5, data for volcanics of the Oyasai, Kyzylnura, Karabash, Chilten, and Babaitaudor calderas respectively; 6–8, data for volcanics of the Tashkesken, Adrasman, and Lashkerek basins; 9, data for volcanics of the Kumlai-Koksarek trough.

DISCUSSION OF THE RADIOMETRIC DATING RESULTS

Reliability of K–Ar dating. The K–Ar age data used in this work have been obtained in labs that apply precision techniques to determine the K and Ar concentrations and use the appropriate instruments and control procedures. Consequently, we can hardly call into question the analytical reliability of the data as a whole. For the investigated volcanics this proves that K–Ar datings corresponding to Early Permian did not result from analytical errors as twenty four radiometric laboratories of the former USSR carried out independent K–Ar dating of one of the liparite whole-rock sample of the Babaitaudor massif. Their results averaged 265 ± 9.6 Ma (Afanas'ev and Zykov, 1970), which corresponds to Early Permian as well (Harland *et al.*, 1990).

However, it is common knowledge that deviations of K–Ar data from the true rock ages can arise not only from analytical errors. The later disturbance of the original K–Ar system, and extrinsic Ar captured or inherited during the rock origin can also lead to a deviation (Faure, 1989). The experience of geochronology shows that examples of such deviation are not uncommon.

To evaluate the reliability of our and previous K–Ar datings we applied a number of methods accepted in geochronological practice.

(1) We compared the K–Ar data for the same sample of volcanic rocks, obtained from different constituents such as hyaline (or micrograined) matrix and phenocrysts. It turns out that dating of this sort yields discordant results in almost all cases (Fig. 5). It is apparent that at least one of the datings of each pair (matrix and phenocryst) does not reflect the true age of volcanism.

Table 2. New data on the K–Ar age on acidic volcanics and other spatially conjugated rocks of the Chatkal–Kurama zone

Ordinal no.	Sample no.	Analyzed material	Concentration		Age, Ma ±2σ
			K, wt %	$^{40}Ar_{rad}$, ng/g	
		Kyzylnura caldera			
		Central subvolcanic massif			
1	30/85	Transparent sanidine	7.86	172.0	291 ± 9
2	30/85	Brick-red orthoclase	7.32	145.9	267 ± 8
3	30/85	Micrograined brown matrix	4.93	92.0	251 ± 8
4	32/85	Brick-red orthoclase	8.03	159.1	265 ± 8
		Chilten caldera			
		Chilten stock of quartz sienite-porphyry			
1	45/90	Biotite	6.43	143.7	296 ± 9
		Babaitaudor caldera			
		Metasomatites of the Nugarzan tin-polymetal ore deposit			
1	58/91	Muscovite	7.94	171.0	287 ± 8
2	63/91	Sericite	5.95	122.5	275 ± 8
		Malikotan dike (Muzbel'-type granite-porphyry)			
1	41/86	Brick-red orthoclase	10.45	209.4	268 ± 8
2	41/86	Micrograined brown matrix	4.42	84.8	257 ± 8
3	65/86	Brick-red orthoclase	9.00	179.0	266 ± 8
		Babaitaudor subvolcanic massif			
1	17/86	Semi-transparent sanidine	7.64	154.5	270 ± 8
2	17/86	Micrograined pink matrix	4.66	94.2	270 ± 8
3	115/90	Micrograined black matrix	4.63	97.3	280 ± 9
		Tashkesken basin			
		Samgar extrusion			
1	B-3	Brown orthoclase	6.04	119.0	264 ± 8
2	B-4	"	8.20	174.0	283 ± 9
3	B-5	"	6.25	122.0	262 ± 8
4	B-9	"	10.00	205.0	274 ± 8
5	B-8 + 9	Biotitic concentrate	5.89	129.0	291 ± 9
6	B-17	Biotite	6.00	131.0	290 ± 9
7	B-17	Brown orthoclase	6.25	116.0	250 ± 8
8	B-19	Pegmatoid alaskite	4.55	88.3	260 ± 10
9	B-19	Pink orthoclase	9.50	197.0	277 ± 8
10	1-G*	Pink orthoclase (concentrate)	5.40	109.7	271 ± 8
11	7-G*	Pink orthoclase	8.12	158.8	262 ± 8
12	8-G*	Pink orthoclase (concentrate)	5.60	107.6	258 ± 8
13	104GT*	"	4.68	85.3	245 ± 10
14	4-OB*	"	5.70	116.6	269 ± 10
15	10-OB*	"	5.65	108.6	258 ± 10
16	M-116*	Trachyliparite	6.28	117.3	251 ± 10
17	M-200	"	4.23	84.4	267 ± 9
18	M-201	"	6.51	123.0	254 ± 10
19	MS-3*	"	6.28	129.0	274 ± 10
20	MS-4*	"	4.92	96.2	262 ± 10

Table 2. (Contd.)

Ordinal no.	Sample no.	Analyzed material	Concentration		Age, Ma ±2σ
			K, wt %	$^{40}Ar_{rad}$, ng/g	
21	4-ST*	Alaskite	5.64	106.2	253 ± 10
22	8-ST*	Pink orthoclase (concentrate)	5.85	110.3	253 ± 10
23	74GT5*	"	5.46	99.1	244 ± 10
24	74-1*	Brown orthoclase	7.73	146.7	255 ± 10
25	36/85	"	7.24	145.6	269 ± 8
26	38/85	"	8.16	158.0	260 ± 8
27	39/85	"	6.16	131.0	283 ± 9
28	39/85	Pink micrograined matrix	5.40	106.0	263 ± 9
29	BN	Brown orthoclase	7.63	160.6	279 ± 10
		Tashkesken Member ("obsidian" horizon)			
1	U-2	Transparent sanidine	7.28	158.0	288 ± 9
2	U-2	Hydrated volcanic glass	3.18	64.3	270 ± 8
		Adrasman basin			
		Kanimansura (nos. 1–11) and Zambaraka (nos. 12–15) ore bodies			
1	10410	Sericite	7.51	144.6	258 ± 8
2	10411	"	8.00	171.5	285 ± 9
3	10412	"	7.53	161.0	285 ± 9
4	10413	"	7.72	161.5	279 ± 9
5	10414	Adularia	7.42	137.5	249 ± 8
6	4176	"	5.85	124.5	283 ± 9
7	4171	Sericite	5.91	126.5	285 ± 10
8	4173	"	7.72	160.0	277 ± 9
9	4174	"	4.47	97.2	289 ± 11
10	4169	"	7.08	153.0	287 ± 10
11	3476	"	7.70	168.0	290 ± 9
12	4170	"	5.87	133.2	301 ± 10
13	4172	"	7.25	161.0	295 ± 10
14	4175	"	6.56	138.5	281 ± 9
15	4177	"	6.08	129.5	284 ± 10
		Sarytaipan Member			
1	U-3	Trachyliparite	7.72	140.2	244 ± 8
2	U-9	Micrograined pink matrix	7.10	141.0	266 ± 8
3	U-9	K-feldspar sphaerolite	11.74	233.0	266 ± 8
		Tavak Member			
1	34/85	Semi-transparent orthoclase	12.45	271.0	289 ± 9
2	34/85	Micrograined pink matrix	5.77	108.0	251 ± 8
3	B-2	Pink orthoclase	12.54	226.0	242 ± 8
4	424-G*	Trachyliparite	4.98	110.0	293 ± 12
		Sphaerolite Member			
1	K-18	Trachyliparite	9.37	192.5	274 ± 8
2	K-15	Liparite	7.22	132.0	246 ± 8
		Lashkerek basin			
		Dike of essexitic porphyry			
1	20/86	Essexite porphyry	2.16	44.8	277 ± 10

Table 2. (Contd.)

Ordinal no.	Sample no.	Analyzed material	Concentration		Age, Ma ±2σ
			K, wt %	$^{40}Ar_{rad}$, ng/g	
		Lashkerek Ignimbritic Member			
1	11/86	Brown orthoclase	12.57	255.0	271 ± 8
2	12/86	"	10.00	203.0	271 ± 8
3	14/86	"	10.36	201.0	260 ± 8
4	851/83	"	8.24	173.1	280 ± 9
5	11/86	Micrograined brown matrix	3.02	59.0	262 ± 9
6	12/86	"	4.00	77.9	261 ± 8
7	14/86	"	4.65	86.0	249 ± 8
		Trachydacitic Member of Nadak Formation			
1	9/85	Transparent sanidine	9.98	211.0	282 ± 9
2	9/85	Micrograined matrix	4.08	76.0	250 ± 8
3	6/85	"	2.97	55.3	250 ± 9
4	6/85	Biotite	7.45	170.5	303 ± 9
5	131/83	"	6.93	158.5	303 ± 9
6	9/85	Chloritized biotite	4.48	99.4	294 ± 9
7	22/85	"	3.92	85.8	291 ± 10
8	19/85	"	2.62	55.0	280 ± 10
9	21/85	"	2.08	41.2	265 ± 12
		Andesitic sequence of Akchasai Formation			
1	572/83	Biotite	6.75	156.6	307 ± 9
		Kumlai-Koksarek trough			
		Kuchar Member			
1	K-1	Biotite	6.84	148.5	289 ± 8
2	K-2	"	6.70	147.2	292 ± 9
3	K-2	Hydrated volcanic glass	1.59	9.4	83 ± 6
		Liparite tuffs			
1	100/90	Micrograined pink matrix	5.00	106.0	282 ± 8

Note: The K–Ar ages of geological bodies in every volcanotectonic structure of the region are placed in the table in ascending order (according to geological data). The samples marked with an asterisk have been dated in the laboratory of radiometric geochronology at the Institute of Geology, Academy of Sciences of Tadzhikistan; other samples – in the laboratory of radiometric geochemistry and geochronology at the Institute of Geology of Ore Deposits, Petrography, Mineralogy and Geochemistry, Russian Academy of Sciences. The samples of ore-bearing metasomatic rocks from Kanimansur and Zambarak ore deposits were donated by V. L. Rusinov (ordinal nos. 1–5), M. L. Lur'e (nos. 6–10 and 12–15), and N. P. Laverov (no. 11). The ore-bearing metasomatic rocks from Naugarzan ore deposit are from the collection of Rusinov. Other samples have been collected by Volkov and Tadzibaev.

(2) We compared K–Ar data of volcanics, for which the relative age relations are well determined from geological observations. Figure 3 shows that K–Ar data almost never display a correct order of stratification of volcanic bodies obtained from geological observations. Consequently, most of the obtained data do not reflect the time of origin of those bodies.

(3) We analyzed the possible correlation between the K–Ar age values for a certain volcanic body and the degree of metamorphism of the corresponding rocks (or minerals). This showed that the dating of one or another sample of volcanic rock (or sample series known to be of the same geological age) yields the minimum ages for mineral fractions with clear evidence of low-temperature hydrothermal metamorphism and the maximum ages for the least altered rock constituents.

It is appropriate to note at this point that all volcanics known in the region demonstrate evidence of this type of metamorphism. The metamorphism was regional in nature and gave rise to: partial or complete substitution of magnetite by iron hydroxides; transformation of water-transparent sanidine into pink, brick-red, or brown perthitic orthoclase partly replaced by pelitic aggregate; albitization and saussuritization of

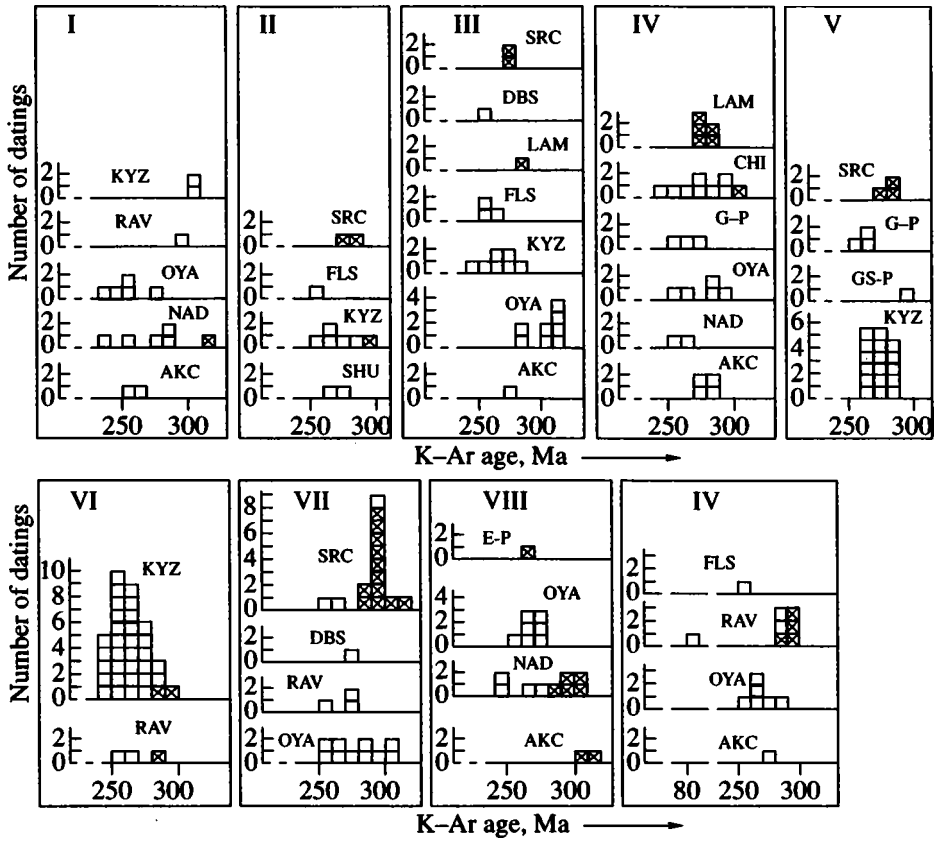

Fig. 3. Correlation between K–Ar ages for magmatic and hydrothermal-metasomatic bodies of the region, whose geological succession is established by field observations: I–V, in the Oyasai, Kyzylnura, Karabash, Chilten and Babaitaudor calderas respectively; VI–VIII, in the Tashkesken, Adrasman and Lashkerek basins; IX, in the Kumlai-Koksarek trough. The formations are labelled on the diagrams as follows: Akchasai (AKC), Nadaksai (NAD), Oyasai (OYA), Shurabsai (SHU), Ravash (RAV), Kyzylnura (KYZ); the Babaitaudor-type granosyenite–porphyry (GS-P), the Muzbel-type granite-porphyry (G-P), the Chilten Complex (CHI), felsite and microgranite dikes (FLS), lamprophyre dikes (LAM), diabase dikes (DBS), ore-bearing sericitoliths (SRC), postmineral dikes of essexite porphyry (E-P). Squares with diagonal cross corespond to minerals, the K–Ar system of which is resistant against the low-temperature metamorphism.

plagioclase; chloritization of biotite and amphibole; local carbonatization of micrograined matrix; and hydration of volcanic glass. The metamorphosed volcanics exhibit pink, red, or brown color. The intensity and completeness of metamorphic alterations vary not only from one outcrop to another, but within the same sample or even the same crystal as well. The petrographic study shows that the rock matrix is always metamorphosed more strongly than phenocrysts, while the described alterations of the phenocrysts develop from their rims toward the cores along cleavage cracks and other structural defects.

An interrelation between the K–Ar data and the degree of metamorphism of the analyzed material can be exemplified by variable K–Ar ages of liparite samples from the Kyzylnura caldera. The age was determined for three rock constituents, which have undergone hydrothermal alteration to a varying degree. The phenocrysts of transparent sanidine show the oldest age (291 ± 9 Ma); the phenocrysts of brick-red orthoclase (metamorphosed sanidine) show a lower value (267 ± 8 Ma); and the minimum age (251 ± 8 Ma) was measured on the micrograined liparite matrix, where K-feldspar is further metamorphosed to pelitic aggregate and colored by iron hydroxides (Table 3, Sample 30/85).

Previously one of us also noted that in the rocks of the Samgar extrusion the K–Ar age of K-feldspar phenocrysts tends to decrease as the structural ordering of this mineral increases (Tadzhibaev, 1979).

Consistently higher K–Ar age values for phenocrysts of the analyzed rocks relative to those for the matrices of the same samples (Fig. 5) can also be related to the variation in degree of metamorphism. This is evident from the fact that, as noted above, the matrix material of all the investigated samples shows a stronger hydrothermal alteration than the phenocrysts.

The variations in K–Ar age of the older Nadaksai Formation volcanics are also significant. The age was determined on biotite phenocrysts chloritized to a varying extent, and as the degree of chloritization increases, the age value decreases from 303 ± 9 to 265 ± 12 Ma (Fig. 6).

Table 3. The results of Rb and Sr isotope analyses for acidic volcanics of the Chatkal–Kurama zone

Ordinal no.	Sample no.	Analyzed material	Concentration, ppm		Atom ratios	
			^{87}Rb	^{86}Sr	$^{87}Rb/^{86}Sr \pm 2\sigma$	$^{87}Sr/^{86}Sr \pm 2\sigma$
		Rocks and minerals of the Babaitaudor massif				
1	55/90	Transparent plagioclase	5	30	0.166 ± 2	0.70696 ± 8
2	55/90	Transparent sanidine	92	8.6	10.61 ± 4	0.74908 ± 5
3	55/90	Granite-porphyry	67	5.8	11.19 ± 6	0.75154 ± 16
4	17/86	Semi-transparent sanidine	78	3.3	23.08 ± 12	0.79855 ± 4
5	18/86	"	81	2.0	40.9 ± 4	0.87082 ± 16
6	17/86	Brown orthoclase	140	2.7	50.2 ± 2	0.90972 ± 10
7	18/86	"	130	1.9	67.2 ± 8	0.9801 ± 3
8	17/86	Liparite	105	1.4	76.9 ± 3	1.0178 ± 2
9	17/86	Micrograined matrix	400	13	90.3 ± 10	1.0759 ± 2
10	18/86	Liparite	135	1.1	122.4 ± 14	1.2054 ± 2
11	18/86	Micrograined matrix	140	1.0	140.2 ± 16	1.2750 ± 2
12	37/90	Liparite	130	0.9	144.8 ± 8	1.2885 ± 2
		$T = 284 \pm 1$ (2σ) Ma; $(^{87}Sr/^{86}Sr)_0 = 0.70627 \pm 15$ (2σ); MSWD = 3.5				
		Rocks and minerals of the Samgar extrusion				
1	GYu-20	Pegmatite	50	8.8	5.75 ± 18	0.7281 ± 7
2	GYu-19	Alaskite	70	10.0	6.86 ± 21	0.7330 ± 7
3	GYu-18	Alaskite-porphyry	64	5.7	11.28 ± 30	0.7492 ± 7
4	GYu-17	Trachyliparite	110	9.7	11.58 ± 30	0.7506 ± 7
5	GYu-19	Biotite	140	5.5	24.98 ± 75	0.8056 ± 7
6	GYu-17	Biotite	140	5.6	26.19 ± 80	0.8105 ± 7
		$T = 281 \pm 8$ (2σ) Ma; $(^{87}Sr/^{86}Sr)_0 = 0.7051 \pm 12$ (2σ); MSWD = 1				
		Mineral components of the Tashkesken Member vitrophyres				
1	U-2	Transparent plagioclase	4.0	3.5	1.124 ± 6	0.71142 ± 4
2	U-2	Transparent feldspars	15	2.9	5.44 ± 2	0.72844 ± 11
3	U-2	Transparent sanidine A	30	1.9	15.68 ± 8	0.76775 ± 5
4	U-2	Transparent sanidine B	31	1.9	16.51 ± 9	0.77264 ± 10
5	U-2	Hydrated glass	74	0.5	142 ± 2	1.2663 ± 2
		$T = 275 \pm 4$ (2σ) Ma; $(^{87}Sr/^{86}Sr)_0 = 0.7070 \pm 2$ (2σ); MSWD = 15.3				
		Rocks and minerals of the Mailikotan dike				
1	65/86	Brown orthoclase	93	5.9	15.49 ± 6	0.76999 ± 10
2	41/86	Brown orthoclase	94	4.9	19.00 ± 12	0.7824 ± 10
3	65/86	Micrograined matrix	64	3.3	19.02 ± 10	0.78337 ± 8
4	41/86	Granite-porphyry	63	2.8	23.40 ± 10	0.79960 ± 6
5	41/86	Micrograined matrix	220	2.3	27.95 ± 10	0.81733 ± 6
		$T = 268 \pm 5$ (2σ) Ma; $(^{87}Sr/^{86}Sr)_0 = 0.7111 \pm 15$ (2σ); MSWD = 4.5				

Note: The rocks and minerals of the Samgar extrusion are analyzed in the laboratory of radiometric geochronology of the Institute of Geology, Academy of Sciences of Tadzhikistan; other samples – in the laboratory of radiometric geochemistry and geochronology, at the Institute of Geology of Ore Deposits, Petrography, Mineralogy and Geochemistry, Russian Academy of Sciences.

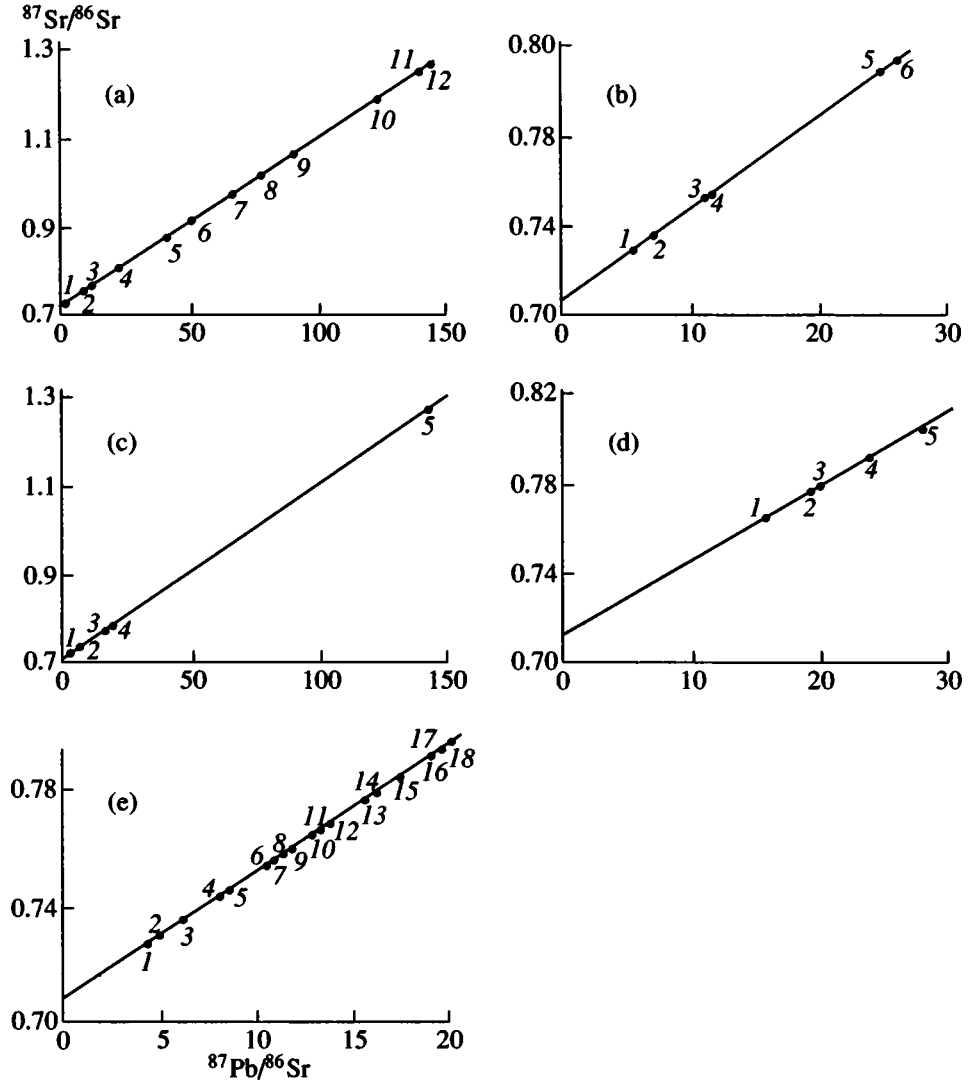

Fig. 4. Rb–Sr isochron diagrams for acidic volcanic bodies of the Chatkal–Kurama zone: Babaitaudor subvolcanic massif (a), Samgar extrusion (b), Tashkesken Member (c), Mailikotan dike (d), and Lashkerek Member (e). The sample numbers on diagrams (a–d) correspond to the ordinal numbers in Table 3 of the present paper and in diagram (e) correspond to those of Volkov *et al.*, 1990 (Table 1).

In summary, the dependence of the K–Ar age values on the degree of low-temperature hydrothermal alteration is evident. This suggests that hydrothermal alteration either directly led to the loss of Ar_{rad} from the volcanic rock phases, or created the prerequisites for such a loss, which occurred later.

4. Some mineral fractions of liparites from the Kyzylnura and Oyasai formations were dated by the ^{40}Ar–^{39}Ar method with the thermal steps of Ar extraction. We intend to report the detail results of this dating in a special publication, although they are already published in brief (Volkov *et al.*, 1987). Essentially, these results show that for transparent sanidine phenocrysts separated from liparite sample 30/86 the value of the Ar–Ar age remains constant (292 Ma) at all temperature steps. Unlike this, the phenocrysts of pelitic brown-colored orthoclase from the same and other samples demonstrate a sequential increase in age values from 258–269 to 282–288 Ma with increasing temperature of Ar extraction.

In general, the spectrum of Ar–Ar ages for sanidine is of a plateau type, while that for orthoclase is of a "ladder" form. The former is typical of the samples without Ar excess and with an undisturbed K–Ar system; the latter is characteristic of the samples that partly lost Ar (Faure, 1989). Ladder-type spectra are obtained for all micrograined matrix fractions from volcanics of Kyzylnura and Oyasai formations, which were analyzed by the Ar–Ar method. These data support the conclusion that the mineral constituents of the investigated volcanics that underwent a low-temperature hydrothermal metamorphism partly lost radiogenic Ar. The data also show that the minerals without any evi-

dence of such metamorphism well retain the radiogenic Ar.

(5) We compared the K–Ar and Rb–Sr ages obtained for the same acidic volcanic bodies of the region (Fig. 7). The comparison shows that in almost all cases the K–Ar datings of the mineral constituents and whole-rock samples with evidences of low-temperature metamorphism yield younger ages than the Rb–Sr isochron method. However, the K–Ar age of transparent sanidine and nonchloritized biotite phenocrysts either is consistent with the Rb–Sr age (within the limits of analytical error) or even exceeds the latter in magnitude. These relationships also support the conclusion on the preservation of Ar_{rad} in minerals unaffected by hydrothermal alteration and on partial loss of Ar_{rad} from minerals that experienced such alteration.

(6) We attach particular significance to a possible shift in K–Ar ages toward older values, which can result from captured or inherited extrinsic Ar. We consider this possibility only for data obtained for minerals lacking evidence of hydrothermal alteration (phenocrysts of sanidine or fresh biotite) for the two following reasons. The data reported above show that hydrothermal metamorphism leads not to the capture, but to the loss of Ar, and only unaltered minerals of volcanics exhibit the oldest ages.

The K–Ar data on such phenocrysts can in fact depart from the genuine age of volcanism toward the older values, if the mineral phases are xenogenous in origin or during crystallization they captured Ar from the magma with a higher ^{40}Ar–^{39}Ar ratio than that in the atmosphere.

However, the assumption of the xenogenous nature of the investigated minerals is in conflict with some geological and petrologic observations (e.g., they show an increase in the number and dimension of these phenocrysts in the central part of volcanic bodies as compared to chilled marginal rocks facies). Even though the minerals are real xenocrysts, the probability that they preserve the previously accumulated radiogenic argon seems very doubtful, because the high temperature of the surrounding melt should cause an intensive Ar diffusion out of the crystals.

We also rule out the possibility of any significant overstating of the K–Ar ages obtained on sanidines and fresh biotites resulting from the capture of excessive Ar dissolved in the magma. In the opinion of geochronologists, who studied the problem of Ar excess in volcanics, any substantial overstating of the K–Ar ages for phenocrysts and glass through Ar capture is possible only for Cenozoic samples, in which the content of radiogenic Ar is low (Damon *et al.*, 1967; Dalrymple and Lanphere, 1969). Calculations show that for minerals like sanidine and biotite, the overstating of K–Ar ages for a value exceeding the analytical error (8–10 Ma for this case) is possible only when the Ar excess in the minerals is as great as $n \times 10^{-10}$ mol/g. However, this is two to three orders greater than Ar contents that were in practice determined in phenocrysts of volcanics and regarded as argon captured from magma (Damon *et al.*, 1967; Dalrymple, 1968a, 1968b; Dalrymple and Lanphere, 1969; Funkhouser *et al.*, 1968). Furthermore, as mentioned above, the spectrum of Ar extraction for the ^{40}Ar–^{39}Ar dating of sanidine phenocrysts (liparite sample 30/85) is of the plateau type. This demonstrates homogenous Ar distribution in the sanidine structure, a feature which is atypical for Ar-excess systems.

In our opinion, the foregoing discussion discredits the assumption of overstated values among the measured K–Ar ages.

Evaluating the reliability of K–Ar ages for the volcanics in question, we may state the following:

1) The K–Ar systems in transparent sanidine and unaltered biotite phenocrysts lack any significant Ar excess and remaining closed during low-temperature metamorphism. Thus, we believe that the K–Ar ages of these minerals yields reliable information on the age of the corresponding volcanic events.

2) The K–Ar systems in altered phenocrysts (orthoclase replaced by pelitic aggregates and chloritized biotite), micrograined to hyaline matrix, and whole-rock samples are disturbed in most cases and show the Ar loss. Accordingly, virtually all K–Ar ages for samples of this kind should be regarded as "rejuvenated."[2] In one way or another, the distortion undoubtedly results from low-temperature hydrothermal metamorphism of the volcanic rocks.

Reliability of the Rb–Sr age data. It has been known that in one or another set of rocks the $^{87}Rb/^{86}Sr$ and $^{87}Sr/^{86}Sr$ ratios can exhibit a linear dependence not only in cases when the Rb–Sr system meets the requirements of the isochron model, but also in cases when the studied samples represent a mixture of two components different in $^{87}Sr/^{86}Sr$ and Rb/Sr ratios (Gorokhov, 1985a), and dating based on the mixture line parameters does not correspond to age indicating a certain geological event.

However, even if the plotted straight line is a true isochron, its age does not always reflect the time of origin of the investigated geological body. Specifically, metamorphism of the rocks could reset the Rb–Sr isotopic geochronometer. In this case the isochron reflects the age of the metamorphic process. Gorokhov (1985b), Rublev *et al.* (1991), Hradetzky and Lippolt (1993) showed that the resetting of the Rb–Sr geochronometers occurs very often in nature and is most pronounced in the case of hydrothermal metamorphism.

In addition, the isochrons based on the whole-rock samples of volcanics can fall into the category of so-called "mantle" isochrons (Gorokhov, 1985b). It is known that ages calculated in this case represent the time of isotope homogenization of the substance in the

[2] Previously Tadzhibaev (1979) and Laverov *et al.* (1985) came to a similar conclusion. However, the data to which they referred provided, in our opinion, for an alternative interpretation as well.

magma source and thus does not reflect the age of volcanism.

The interpretation of linear relationships (Fig. 4) is acceptable as valid if it conforms to the peculiarities of the analyzed samples and does not conflict with the geological succession of the dated bodies.

It is significant for the interpretation that, as noted above, the straight lines in Fig. 4 approximate the data points not only for whole-rock samples, but also for the mineral and other rock fractions that crystallized from the magma (phenocrysts etc.). This does not allow us to interpret these straight lines as "mantle" isochrons, because data for magmatic minerals cannot fall on such isochrons. Indeed, the Rb–Sr systems of magmatic minerals are initiated only at the moment of magma crystallization and thus cannot bear any information on the $^{87}Sr/^{86}Sr$ and $^{87}Rb/^{86}Sr$ ratios in the source material at the time of its homogenization.

We also have to exclude the assumption that the straight lines in our diagrams are mixing lines.

Evidently the mixing model can be realized for volcanic rocks in the following two ways: through contamination of the primary magma by an extrinsic material, and through chemical alteration of already formed volcanics under the influence of hydrothermal solutions. Petrographic observation supports the involvement of contamination and hydrothermal alteration in creating the geochemical peculiarities of the volcanics under study.

However, we cannot use the model of melt contamination to interpret the straight lines in Fig. 4, because the mixing of melt and contaminating material controls parameters of the Rb–Sr system only in the whole-rock samples. As for magmatic minerals in these rocks, their contents of Rb and Sr, as well as the contents of all trace elements, generally depends on the crystal–melt distribution coefficients. Thus it is evident that data points for magmatic minerals can fall only accidentally on the mixing line approximating whole-rock compositions. As for our straight lines, they regularly fit both the whole-rock and mineral compositions, and this cannot be interpreted as an accident.

In regard to the mixing between the volcanic rocks and hydrothermal solutions, it is evident that in the case under consideration there was a mixing of more than two components. Indeed, the rocks of different phases and facies of any volcanic body, as well as their constituent minerals, undoubtedly varied in $^{87}Rb/^{86}Sr$ ratio (and also most likely in $^{87}Sr/^{86}Sr$ ratio) at the moment of metamorphism. Because of this, the hydrothermal metamorphism could not result in linear correlation between the $^{87}Rb/^{86}Sr$ and $^{87}Sr/^{86}Sr$ ratios measured in our sample collection.

Admittedly, the discussed linear relationships are ordinary isochrons. Therefore calculated isochron ages represent either the age of volcanism or the age of superimposed metamorphism.

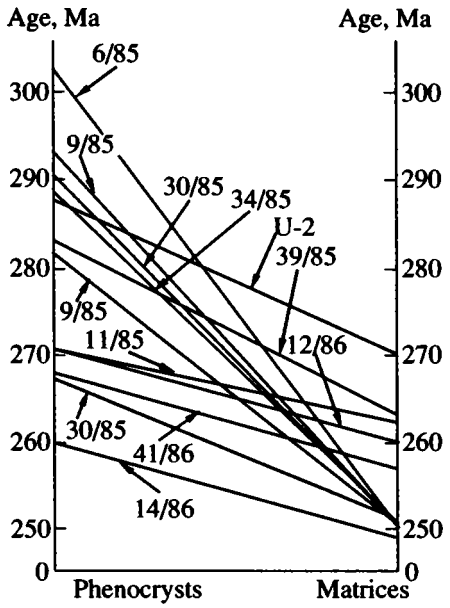

Fig. 5. Correlation between K–Ar ages for Late Paleozoic volcanics of the Chatkal–Kurama zone measured in coexisting (within the same sample) phenocrysts and matrix. The analytical data for the samples pointed by arrows are from Table 2 (see for rows with corresponding numbers).

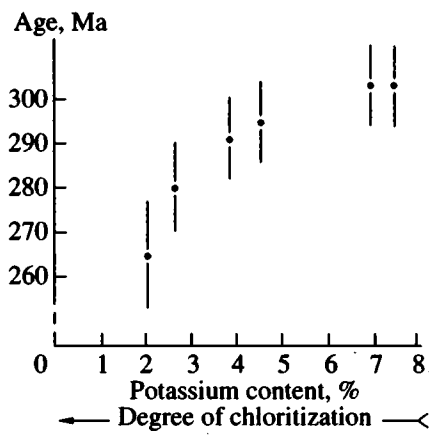

Fig. 6. K–Ar age of biotite versus degree of its chloritization. Analytical data are from Table 2. Vertical lines show ranges of the analytical error (2σ).

In choosing one of these interpretations we must take into account and explain the fact that the isochrons under consideration approximate the data points of the samples, whose material is altered to a widely varying degree by hydrothermal metamorphism.

Since each isochron combines compositional points of rocks and minerals with a distinct degree of metamorphism, the following three situations are possible: (1) the metamorphism was almost synchronous with the volcanism; (2) the metamorphism occurred much later than the volcanism and completely changed the Rb–Sr systems of all constituents in volcanic rock including those without visible evidences of metamor-

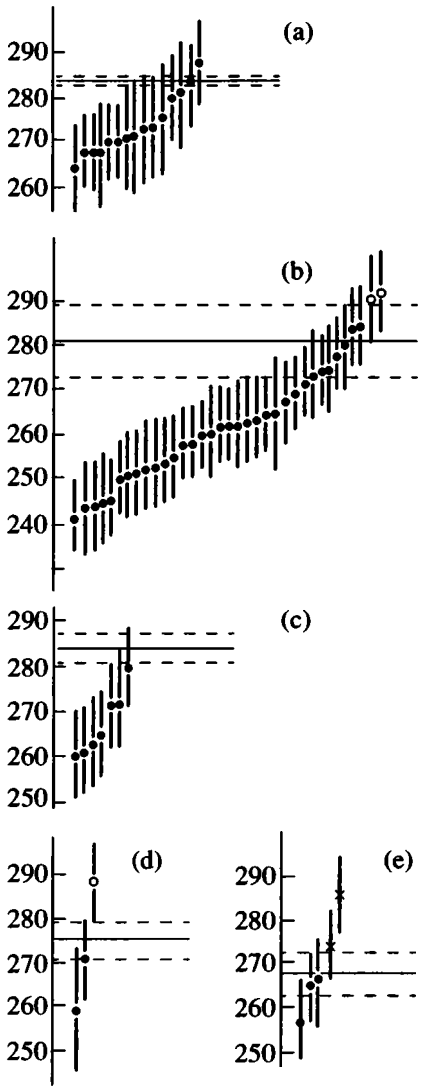

Fig. 7. Comparison of Rb–Sr isochron and K–Ar ages (the latter arranged by their values) for the Babaitaudor massif (a), Samgar extrusion (b), Lashkerek Member (c), Tashkesken Member (d), and Mailikotan dike (e). Horizontal lines mark levels of Rb–Sr ages; points show K–Ar ages of volcanic rock fractions with clear evidences of low-temperature hydrothermal alterations; circles indicate K–Ar ages of the minerals without such evidences; crosses are for K–Ar ages of muscovite and sericite from metasomatic rocks substituting granite porphyries in Mailikotan dike. The horizontal dashed and vertical solid lines deliniate analytical errors (2σ) of Rb–Sr and K–Ar data respectively.

phic alterations; (3) the metamorphism was separated in time from the volcanism and did not disturb the Rb–Sr systems even in the minerals with clear evidences of hydrothermal alteration. The first or second situation means that the isochron ages reflect the age of metamorphism. In the third case the dating should represent the age of volcanism.

It is beyond question that Rb–Sr systems of the analyzed volcanics did not remain closed during metamorphism. This is proved by a marked increase of Rb content in brown pelitic orthoclase formed after sanidine during the metamorphism of volcanics (see data on sanidines and orthoclases, samples nos. 17/86 and 18/86, Table 3).

Therefore the isochrons under examination could correspond only to either the first or second situations listed above. Consequently, the obtained isochron ages reflect an age of metamorphism that was either synchronous to (the first case) or separated in time from the volcanism (the second case). This idea conforms well with the character of relations between the discussed Rb–Sr and K–Ar ages obtained for the same volcanic bodies. The latter ages were measured on minerals without evidence of low-temperature alterations, such as sanidine and fresh biotite phenocrysts, and also on muscovite and sericite produced by high-temperature metasomatism affected by the volcanics (Figs. 7b, 7d, and 7e). In all comparative analyses, the value of K–Ar age is always higher than that of Rb–Sr age and sometimes the difference is beyond the limits of analytical error.

Rublev *et al.* (1991) noted analogous relationships between K–Ar and isochron Rb–Sr ages obtained on phenocrysts and whole-rock volcanic samples from Siberia respectively. Using independent methods, these authors showed that the K–Ar data correspond to the age of volcanism, while the Rb–Sr determinations reflect the age of low-temperature hydrothermal metamorphism, the evidence for which is visible in the rocks.

The similarity between these and our results suggests that the Rb–Sr isochron ages in the studied region also correspond to the time of low-temperature metamorphism.

Thus, the measured isotropic ages of acidic volcanics from the Chatkal–Kurama zone shows that neither the K–Ar, nor the Rb–Sr values obtained can be regarded as overstated relative to the age of a corresponding volcanic event. On the contrary, in most cases we have to consider the K–Ar ages to be relatively rejuvenated. As is shown, the results of Rb–Sr isochron dating reflect the age of metamorphism, which affected the rocks at the time of volcanism or afterwards. Thus the obtained Rb–Sr isochron data represent the age of volcanism only in some particular cases. The K–Ar ages for transparent sanidine and unaltered biotite phenocrysts represent the most reliable data on the time of volcanic events.

CONCLUSION

Taking into account the isotopic age determination and estimated validity of obtained values, we draw the following conclusions about the timing of hyperacidic volcanism in the region.

(1) The available isotopic ages substantiate only the minimum age limit for volcanic bodies traditionally

regarded as members of the Oyasai Formation. For most of them this age corresponds to the Asselian Stage of the Early Permian. Since all fossil floras, which occur in the rocks assigned to the Oyasai Formation, are of the Early Permian age, and one rock member yields Asselian fusulinids, it is safe to say that radiometric and paleontological ages of the formation are concordant and suggest that most volcanic bodies of this formation were formed in the Asselian.

However, we should emphasize that the K–Ar dating of hyperacidic volcanics of the Chilten and Karabash calderas, which are also assigned to the Oyasai Formation, yielded several values corresponding to the Middle Carboniferous (Fig. 3). This permits an assumption that some of the hyperacidic volcanic bodies within the mentioned calderas can be older than the Asselian (see below).

(2) Judging from the K–Ar ages on the most reliable mineral geochronometers (288 ± 9 Ma on sanidine from an "obsidian" of the Tashkesken Member; 289 ± 8, 292 ± 9, 295 ± 9, and 297 ± 12 Ma on fresh biotites from trachydacites of the Kuchar Member), the Ravash Formation was formed in the Asselian Stage, in other words, synchronously with the Oyasai Formation.

However, the age of the Ravash Formation cannot yet be considered as absolutely reliable at present, because the Tashkesken and Kuchar rocks dated radiometrically are not generally accepted as its members.

(3) All the largest volcanic bodies assigned by previous researchers to the Kyzylnura Formation were formed 285–290 Ma ago, i.e., in the Asselian Stage of the Early Permian. This is supported by the following data:

(a) K–Ar ages of 291 ± 9, 290 ± 9, and 291 ± 8 Ma are obtained respectively on sanidine phenocrysts from rocks of the central subvolcanic massif of the Kyzylnura caldera, on two fresh biotites from the Samgar extrusion, and on muscovite from greisen replacing granite-porphyries of the Mailikotan dike (Table 2);

(b) Rb–Sr isochron ages of 284 ± 1.5 and 281 ± 8 Ma were measured in the Babaitaudor massif and Samgar extrusion (Table 3).

(4) The isotopic age of the investigated stratotype deposits of the Kyzylnura Formation coincides with the age of the formation determined by paleontologic methods (Arapov *et al.*, 1978) and our radiometric data. This fact, as well as the compositional similarity between the Kyzylnura and Oyasai volcanics and their coexistence in the same structures, make us agree with the idea of Dalimov and Rafikov that geological bodies of both formations represent the same unit in many, or even in most cases.

(5) However, contrary to Dalimov and Rafikov, we do not yet rule out a possibility that two acidic volcanic formations of different age are present in the region, because some geochronological data support this possibility. In particular, acidic volcanics of the Karabash depression, which were assigned to the Oyasai Formation, yielded unusually old (Middle Carboniferous) K–Ar ages of 312, 311, and 307 Ma even on whole-rock samples (*Katalog opredelenii ...*, 1972). Clearly, rocks with such age marks cannot be assigned to the Asselian. The felsoliparite sequence enclosing the Chilten stock (Chilten caldera) seems to be also a member of the older formation, since we obtained a K–Ar age of 296 ± 9 Ma on the rocks of this stock (Table 3).

(6) The isotopic data do not support the idea that some acidic volcanics younger than Asselian are present in the region. Of course, the hypothesis about the presence of such rocks has the right to exist until the opposite is proved for absolutely all acidic volcanic bodies of the region. However, up to now, all isotopic-geochronological and geological data convincingly show that if a later event of acidic volcanism did occur in the region, the scale of its manifestation was many times smaller than that documented in the Asselian Stage.

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International Symposium: "Evolution of Permian Marine Biota," and Problems Concerning Permian Sections in the Type Region

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Abstract—The main results of the International Symposium "Evolution of Permian Marine Biota," are presented. The Artinskian to Ufimian part of the section at the Kozhim River (Polar Urals) is briefly described including its paleontological characteristics. Potentials for further study of Permian deposits in the Urals and possibilities of using the Uralian stages as international standards are discussed.

Key words: *Artinskian Stage, Kungurian Stage, Ufimian Stage, ammonoids, brachiopods, bryozoans, stratotype, parastratotype*

The International Symposium: "Evolution of Permian Marine Biota," with a field excursion to Permian deposits at the Kozhim River, was organized by the Paleontological Institute of the Russian Academy of Sciences and held August 15–25, 1995. Twenty participants from various Russian institutes such as the Paleontological Institute, the Geological Institute (Moscow, Russian Academy of Sciences), the Komi Geological Institute (Syktyvkar, Ural Division of the Russian Academy of Sciences), Perm' State University, Saratov State University, the All-Russian Research Institute of Geological Prospecting (Sankt-Peterburg), and one participant from the University of Melbourne (Australia) were on the field excursion on August 15–20. Fifty participants from Russia, three participants from the Nanking Geological Institute (China) and one participant from the University of Melbourne (Australia) attended the general session. Over 30 oral presentations focused on various problems of the evolution of Permian marine organisms and their biogeography and stratigraphic importance. Abstracts were published in Russian and English; the guide-book for field excursions to the Permian sequence at the Kozhim River is in Russian.

The participants of the symposium were very interested in the Permian sequence at the Kozhim River (Polar Urals (Figs. 1 and 2)). For one week participants had an opportunity to study this unique sequence and to collect fossils from carbonate and terrigenous sediments spanning the interval from the Upper Carboniferous to the Ufimian Stage of the Upper Permian (Kuz'kokova *et al.*, 1980; *Permskaya sistema* ..., 1991; *Permian system* ..., 1993; *Putevoditel'* ..., 1995). The most interesting is the interval from the Artinskian to Ufimian Stage (Fig. 2). This interval is briefly described below (from the base upward):

(1) The Kos'ya Formation is subdivided into siltstone (lower) and sandstone (upper) subformations.

The siltstone subformation [the Sakmarian (Sterlitamak Horizon) – Artinskian (Burtsevo Horizon)] has a thickness up to 400 m and is represented by siltstones and mudstones with minor intercalations and lenses of detritic limestones.

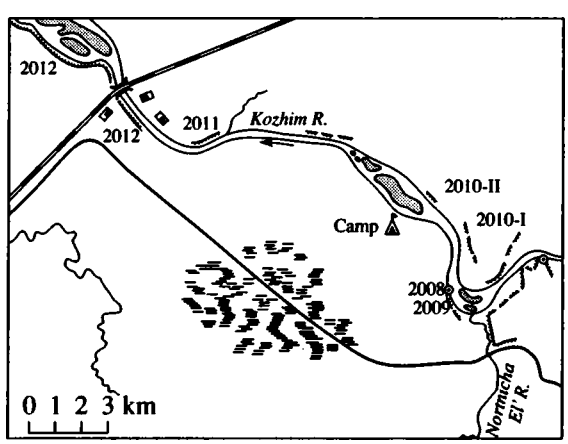

Fig. 1. Map of the Permian outcrops at the Kozhim River.

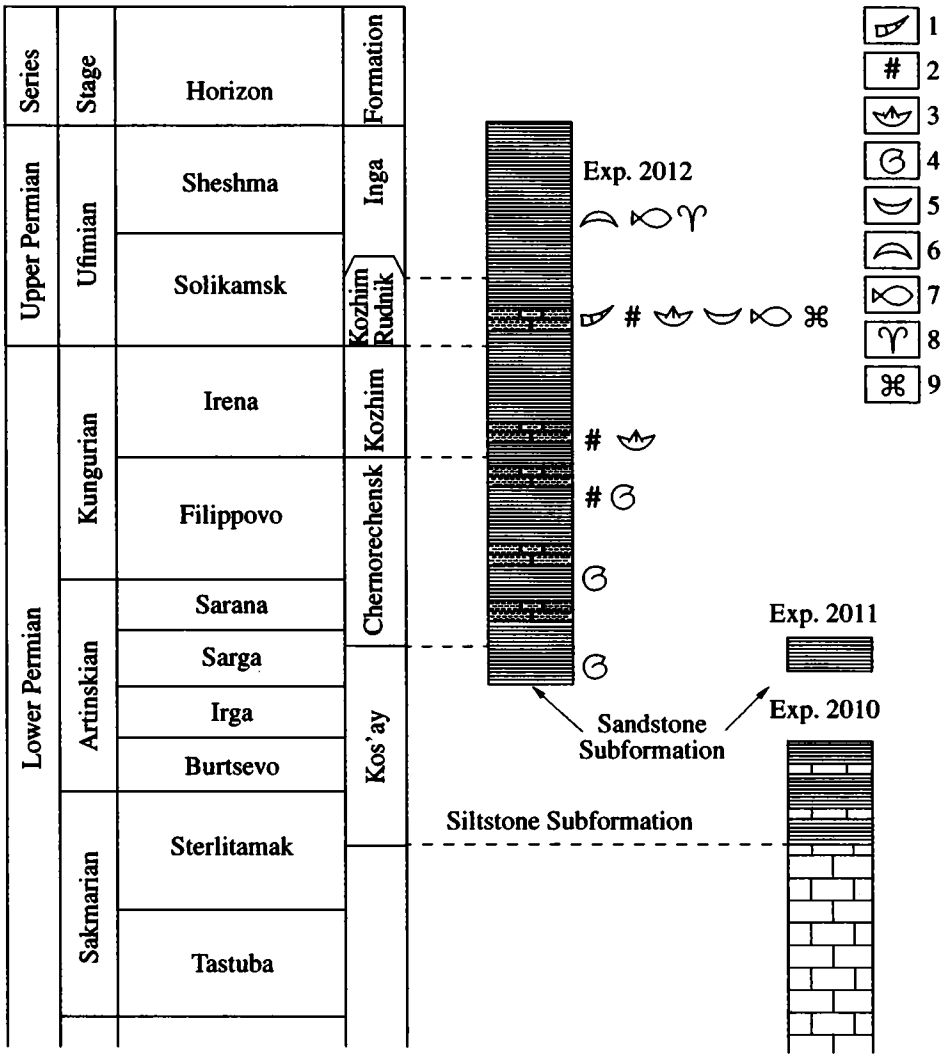

Fig. 2. The Permian sections at the Kozhim River with indicated occurrence levels of: 1, foraminifers; 2, bryozoans; 3, brachiopods; 4, ammonoids; 5, marine bivalves; 6, non-marine bivalves; 7, fishes; 8, tetrapods; 9, flora.

The sandstone subformation (the Artinskian Irgina and Sarga horizons) has a thickness up to 200 m and is represented by massive sandstones with subordinate siltstone beds. The ammonoids *Paragastrioceras jossae exile* Ruzh., *P. kirghizorum* Voin., *Sakmarites vulgaris* (Karp.), *Artinskia ariensis* (Grun.), *Medlicottia orbignyana* (Vern.), *Waagenina subinterrupta* (Krot.), and *Uraloceras suessi* (Karp.) were collected from the middle part of this subformation.

The Chernaya Rechka Formation [the Artinskian (Sarga and Sarana horizons) – Kungurian (Filippovo Horizon)] is represented by thick alternating dark-grey and greenish-grey siltstones with subordinate beds of marls and pelitic limestones. The ammonoids *Paragastrioceras* sp., *Uraloceras* sp. cf. *U. inzolutum* (Voin.), and *Uraloceras* sp., were collected at the level of 395 m above the base of the Chernaya Rechka Formation. The 385-m-thick upper part of the Chernaya Rechka Formation is correlated with the Sarana Horizon of the Artinskian Stage and with the Filippovo Horizon of the Kungurian Stage. The ammonoids *Paragastrioceras* sp., and abundant bryozoans were found in this member. The bryozoans are represented by *Ramiporidra* sp., *Goniocladia* cf. *compacta* Schulga-Nest., *Laxifenestella sublatericrescens* (Schulga-Nest.), *L. mariae* (Trizna), *Alternifenestella pseudobifida* (Schulga-Nest.), *Parseptopora uralica* (Nikif.), and *Polyporella trigonocella* (Schulga-Nest.). According to data on myospores, the upper part of the formation (approximately 140 meters thick) can correspond to the Filippovo Horizon of the Kungurian Stage (*Permian system* ..., 1993).

The Kozhim Formation (the Kungurian Stage, Irena Horizon) is subdivided into two parts. The lower part is 105 m thick and represented by sandstones with minor interlayers and beds of siltstones and mudstones. Rare

sandstone beds of 2–6 m in thickness are present also. Brachiopods *Derbyia, Neochonetes, Krotovia, Kochiproductus, Sowerbina, Anemonaria, Thuleproductus, Waagenoconcha, Kaninospirifer, Spiriferella, Rhynchopora, Stenoscisma, Pinegathyris* and others are the most abundant fossils of the Kozhim Formation. These brachiopods are typical of various facies in nearly all known basins and can be used for detailed correlations with the type section of the Kungurian Stage and with more distant sections of the Canadian Arctic Archipelago, Spitzbergen, Australia, southern and Inner Mongolia. This stratigraphic interval also yielded abundant foraminifers (10 levels), bivalves (7 levels), myospores (11 levels), bryozoans, ostracodes, crinoids, and ganoid fishes. These deposits are undoubtedly Kungurian in age.

The Kozhim Rudnik Formation (the Ufimian Stage, Solikamsk Horizon) is represented by alternation of thin-bedded (2–25 cm thick) shales, siltstones, and sandstones, whose total thickness is approximately 550 m. Sandstones are greenish-grey, fine- to medium-grained with parallel and cross-bedding. They form beds 2–18 m thick and contain thin (0.1–1 m) layers with coalified organic detritus.

The deposits of the Kozhim Rudnik Formation contain abundant fossils (foraminifers, bryozoans, brachiopods, bivalves, fishes, leaved flora, and myospores). The majority of experts on marine invertebrates and myospores determine these deposits as corresponding to the Ufimian Stage, but some paleontologists, especially experts on leaved flora, recognize it as a Kungurian unit.

The Inta Formation (the Ufimian Stage, Solikamsk and Sheshma horizons) is represented by the alternation of relatively thin (0.5–1.5 m thick) layers of sandstones, siltstones, and shales with a total thickness of 580 m. Beds of cross-bedded sandstones (up to 36 m thick) contain abundant myospores and leaved flora. Non-marine bivalves, fishes and tetrapods were also found in this interval as well as some coal seams. These deposits are interpreted to be fluvial facies.

The participants of the symposium indicated that the Kozhim section has a high potential for correlation with stratotypes of the central and southern Urals. After the field excursion and discussion, the following recommendations were adopted.

The standard of the Permian System was originally distinguished among sections of the Russian platform and the Urals and needs additional substantiation by various modern stratigraphic methods. The Asselian, Sakmarian and Artinskian stages are accepted to be its standard units, which can be regarded as constituting a special Cisuralian Series. The Kungurian, Ufimian, Kazanian and Tatarian stages are represented in the stratotype region by lagoonal and continental deposits. Their identification in marine sections is difficult. Alternative scales based on the Tethyan and the North American sections were proposed for this stratigraphic interval. These scales appeared because they do not have the disadvantages peculiar of the aforementioned stages and also because their stratotypes have not been supplemented with sufficient information.

However, data obtained from recent studies show that stratotypes of these Permian stages are of high correlative potential. It is important that these units were established and substantiated within a single large region. By virtue of this fact, mistakes in direct interregional correlation can be excluded. The presence of diverse palynologic assemblages and extremely abundant continental flora and fauna can also be considered advantages of the central Russian and Uralian sections. The correlation potential of the stage boundaries can be significantly improved by additional data on conchostracs, fishes, kharophyts, insects and other still poorly studied fossils, and by using other, e.g. paleomagnetic, statigraphic methods. The correlation potential of some units (especially the Kungurian and Ufimian stages) can be refined by introducing additional marine parastratotypes of the Permian stratotype area.

The Kozhim River section (western slope of the Polar Urals) can be considered an appropriate parastratotype of the Kungurian and Ufimian stages. This section meets the main requirements of the international standards. Nevertheless, additional studies are necessary here to find more convincing evidence that the stage boundaries are synchronous in the stratotype and parastratotype sections. The following decisions were recommended by the symposium participants.

The boundaries of the Kungurian Stage must be properly selected and substituted in the stratotype section. Actually, the lower boundary of the Kungurian Stage can be established at one of the following three levels: either at the base of the Sarana and Filippivo horizons, or at the base of the Nevolon Member of the Irena Horizon. The upper stage boundary can be placed at the base, or at the top of the Solikamsk Horizon. The correspondence of the upper Kungurian boundary to the top of the Solikamsk Horizon seems to be better supported by biostratigraphic and other geologic data.

Additional study of all fossil groups from the parastratotypes of the Kungurian and Ufimian stages is essential. The most important are brachiopods, whose high correlative potential was shown in contributions to the symposium. The respective collections of Z.Z. Gizatulin (Kazan State University), N.V. Kalashnikov (Komi Geological Institute of Russian Academy of Sciences) and M.V. Kulikov (All-Russian Research Institute of Geology) can be used as a base. Experts on marine bivalves, ammonoids and flora should urgently analyze relevant data on the stratotype sections of the Chernaya Rechka, Kozhim and Kozhim Rudnik formations and their analogues. They should also reach a satisfactory decision on how to correlate the boundaries of these units.

Many experts suppose that the Roadian Stage is the most probable analogue of the Ufimian Stage. The joint Russian–American study of the Roadian Stage stratotype and coeval deposits in the United States and Canada will be very important.

The symposium nominated a working group for the scientific organization and coordination of the proposed studies. B.I. Chuvashov (corresponding member of Russian Academy of Sciences) was recommended as the head of the group. M.F. Bogoslovskaya (Paleontological Institute, expert on ammonoids), T.A. Grunt (Paleontological Institute, expert on brachiopods), G.P. Kanev (Komi Geological Institute, expert on bivalves), and S.K. Pukhonto (State Geological Enterprise "Polyarnouralgeologiya," expert on flora) were recommended as members of the group. G.V. Kotlyar (All-Russian Research Institute of Geology, expert on the Far East), E.Ya. Leven (Geological Institute, expert on the Tethys), I.N. Manankov (Paleontological Institute, expert on Mongolia), N. Archbold, and Guan R. Shi (experts on Australia) were recommended as leaders of regional working groups. A special working group for additional study of the Kazanian and Tatarian stages should be organized at Kazan State University under the leadership of N.K. Esaulova and V.G. Halymbadzha.

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